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The antarctic winter; simulations with climatological and reduced sea-ice extents.

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Abstract

The Southern Hemisphere climatology of a recent version of the Meteorological Office 11-layer general circulation model is presented. The simulated climate is compared with observational data, including Meteorological Office operational analyses for the period 1983-1987. The surface temperatures and depth and position of the antarctic circumpolar trough are substantially more realistic than found in earlier climate models, but the high latitude upper troposphere is still colder than observed giving an excessively strong westerly jet. The sensitivity of the model to changes in the distribution of sea-ice is investigated and compared with that found in other models. In a first experiment, all Southern Hemisphere sea-ice equatorward of 67.5°S is replaced by sea at 271.2 K, with a consequent reduction in surface roughness length from 10^{-1}m to 10^{-4} m. In a second experiment only the roughness length was changed. It is found that the change in surface roughness contributes substantially to the response to reduced sea-ice extents. The specification of surface roughness over sea-ice in numerical models is discussed.

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1. Introduction

Much recent research in climate change has involved the use of numerical models (see for example Houghton 1984). The changes in climate simulated by such models are dependent on the simulation of the unperturbed climate (Spelman and Manabe 1984, Palmer and Mansfield 1986, Mitchell et al 1987), indicating that the ability to simulate present day climate accurately is a necessary though not sufficient condition for a model to produce reliable estimates of climate change. Thus, it is important that models used in climate change experiments are validated against the present climate. There is a wealth of atmospheric data for the Northern Hemisphere and low latitudes of the Southern Hemisphere which can be used to verify climate model simulations (for example Oort and Rasmussen 1971, Newell et al, 1972, Shea 1986) and although some climatologies do cover the whole of the Southern Hemisphere (for example Schutz and Gates 1971, 1972) the distribution of observations is much sparser than for the Northern Hemisphere (see Figure 1 of Hecht, 1985).

Over the last decade, the availability of atmospheric data for the Southern Hemisphere has increased as a result of both increased satellite coverage, and the advent of global numerical weather prediction models such as those at the Meteorological Office and at the European Centre for Medium Range Forecasts. Such centres analyse and archive global data on a daily basis. Operational analyses are being collated to provide a valuable source of climatological data although they may contain biases due to shortcomings in both the analysis scheme and in the numerical weather prediction model used to assimilate the observations. Thus information on both the mean state of the atmosphere and its variability are now becoming readily available for the whole of the Southern Hemisphere.

There have been many assessments of the simulation of climate by general circulation models (GCM's), but few have concentrated on southern high latitudes. Herman and Johnson (1980) described the polar climatology of the GLAS model, and briefly reviewed simulations by other contemporary GCMs, and more recently Schlesinger (1984) reviewed simulations of several atmospheric GCMs. The principle deficiencies in the models include an underestimation of the depth and errors in the position of the circumpolar surface pressure trough particularly in low resolution models, and a tendency to produce surface temperatures substantially higher than observed over Antarctica. Herman and Johnson (1980) also report a tendency to underestimate upper tropospheric temperatures in high latitudes of both hemispheres. In the first half of this paper we assess the simulation of a recent version of the Meteorological Office 11-layer model in middle and high latitudes of the Southern Hemisphere in winter, taking advantage of the increased availability of observational data.

During winter, sea-ice covers up to 18 x 10^6 km^2 of the southern hemisphere. Sea-ice substantially reduces the transfer of sensible and latent heat from the ocean to the atmosphere. Observational estimates made in the Arctic of the flux of heat from the surface into the atmosphere (see for example Mitchell and Hills, 1986) increase by at least an order of magnitude as one goes from the central Arctic to the neighbouring ocean. Given that sea-ice anomalies persist for several months (Lemke et al 1980), one might expect the associated anomalies in surface heat flux to influence the atmospheric circulation. Many observational studies (for example Wiese, 1924; Schell, 1970; Ratcliffe and Morris, 1978) and several numerical experiments (Williams et al 1974, Herman and Johnson 1978) have attempted to determine the influence of variations in Arctic sea-ice extents on the overlying circulation. There is some evidence that a decrease in Arctic sea-ice extents is associated with local falls in surface pressure.

Whereas in the Arctic, much sea-ice is perennial, around Antarctica most pack ice melts in summer and reforms in winter. Little is known about the nature of the ice, though evidence from satellite data (for example Zwally et al 1979) indicates that open water is more extensive than in the Arctic. Few direct measurements of surface fluxes have been made, so most observations are estimates based on climatological data (Andreas et al, (1984) report measurements indicating an average <u>downward</u> flux of sensible heat during a traverse of the antarctic marginal ice zone in October, but northerly winds carrying warm maritime air over the ice pack persisted for most of the passage, so conditions were highly anomalous). Estimates of

the upward turbulent heat flux over pack ice made by Weller (1980) and over the open sea by Zillman (1972) and Van Loon (1984) show a much smaller contrast between the pack ice and the neighbouring ocean than found in the Arctic. Thus, one might not expect a strong relationship between antarctic sea-ice extents and the overlying atmospheric circulation. Indeed, observational studies (Ackley 1981; Streten and Pike 1980; Cavalieri and Parkinson, 1981) are generally inconclusive, although this may be due in part to insufficient or poor data.

Several numerical studies have investigated the effects of varying antarctic sea-ice extents (Simmonds, 1981 henceforth referred to as S; Mitchell and Hills, 1986 (MH); Simmonds and Dix, 1987 (SD)). S replaced the September sea-ice distribution with that for March in a perpetual September simulation using a hemispheric model, and found reduced westerly tropospheric flow and both reductions and increases in surface pressure in the region from which sea-ice was removed. MH removed all sea-ice equatorward of 66°S in 3 winter simulations performed with a global atmospheric model, reporting similar changes in tropospheric flow to that noted by S, but a pronounced reduction in surface pressure in the regions in which sea-ice was removed. SD commented on the findings of MH, and presented results obtained using a global version of S's model with an improved formulation of surface exchanges. They removed all Southern Hemisphere sea-ice in a 450 day perpetual July simulation, finding reduced westerly flow in the troposphere as in S and MH, and both rises and falls in surface pressure in the region from which sea-ice was removed, as in S. The geographical distribution of the changes, however, differed from that in S. For further discussion of these experiments the reader is referred

to Simmonds and Dix 1986, SD, and the response by Mitchell and Hills (1987). Note however, that SD point out that in the control simulation used by MH, the circumpolar surface pressure trough is too weak and displaced to the north of its observed position. Hence, in the second part of this paper we describe results from a repeat of MH's experiment using a version of the 11-layer model which produces a control simulation superior to those reported in S, MH or SD in order to establish that MH's findings are not associated with deficiencies in their control simulation.

Mitchell and Hills (1987) note that in both S and SD, the same formulation is used for the surface exchange coefficients over sea-ice as over the ocean whereas in the Meteorological Office models much larger values are used over sea-ice. Banke et al (1980) report that a typical value of the drag coefficient over Arctic pack-ice free of major ridges is 1.6×10^{-3} with a maximum value of 2.3×10^{-3} as opposed to a typical value of 1.3×10^{-3} over the ocean (Large and Pond, 1981). Measurements made in the marginal ice zone (MIZ) generally give higher values due to greater surface roughness. Overland and Walter (1985) derived values of 2.5×10^{-3} to 3.1×10^{-3} using aircraft measurements made over the MIZ in the Bering Sea, and Mackling (1983) estimated a mean drag coefficient of 3.1×10^{-3} from 138 profiles of mean wind speed and temperature taken in the inner MIZ of the Bering Sea. Note that drag coefficent depends primarily on surface roughness but also on static stability. In the cases in the Bering Sea the stability was near neutral; however Smith et al (1970) found higher values over the Gulf of St Lawrence in unstable conditions. Data on antarctic sea-ice is limited due to the extreme difficulty of making measurements. Recently, several experiments have been undertaken in the MIZ around

Antarctica to investigate the atmospheric boundary layer. Andreas et al (1984) deduced a value of 4.0×10^{-3} for the surface drag coefficient in 80% ice cover, close to three times the typical ocean value and twice that reported by Banke et al (1980) for close rough sea-ice. Andreas et al (1984) suggested that the drag coefficient achieves a maximum value at approximately 80% ice concentration and thereafter falls to a value 2 to 3 times smaller over a 100% ice cover. As indicated earlier, evidence derived from satellite data suggests ice in the antarctic is more broken up by areas of open water than in the Arctic; and it may thus be aerodynamically rougher. Bennett and Hunkins (1986), on the basis of results from a two dimensional model, speculate that Andreas et al (1984)'s estimates of the drag coefficient over compact sea-ice are too high, but Andreas (1987) comments that the model results appear to support his original findings when the appropriate sea-ice concentrations and drag coefficients are used. A more comprehensive review of drag coefficients over sea-ice is given by Overland (1985).

Given the observational evidence, it may be more realistic to assume a larger value over antarctic sea-ice than over the ocean, as in the Meteorological Office models. In the third part of this paper, we describe an additional experiment in which the formulation of drag coefficients over all Southern Hemisphere sea-ice equatorward of 67.5°S is set to that appropriate to the ocean as opposed to that for land. This is compared with the control simulation to determine the effect of changing the surface roughness, and with the sea-ice anomaly experiment to isolate the effect of the increase in turbulent heating from the surface.

To summarize, the aims of this paper are:

i. To assess the simulation of the antarctic winter by a recent version of the Meteorological Office 11-layer model.

ii. To repeat Mitchell and Hill's 1986 sea-ice anomaly experiment using the Meteorological Office 11-layer model and to compare the results with the earlier experiment.

iii. To investigate the relative size of the changes in circulation due to changes in turbulent heating and to changes in surface drag as a result of reducing sea-ice extents.

2. The model and its antarctic winter climatology

a. The model

The Meteorological Office 11-layer model is a global finite difference model with a regular 2.5° x 3.75° latitude/longitude grid, and 11 sigma (σ = pressure/surface pressure) layers which are irregularly spaced, being concentrated near the boundary layer and near and above the tropopause. The version of the model is that described in detail by Slingo et al (1988). The seasonal and diurnal variations of solar radiation are represented and long wave fluxes are calculated using an emissivity approximation (Slingo and Wilderspin, 1986); cloud amounts are determined within the model. Sea surface

temperatures and sea-ice extents are prescribed from climatology, and updated every 5 days. A thickness of 2 m is assumed in calculating the heat flux through sea-ice.

The boundary layer may occupy the lowest one to three layers, following an extension by Richards (1980) of the method proposed by Clarke (1970). The calculation of the fluxes of heat, moisture and momentum from the surface is based on Monin-Obukhov similarity theory assuming a roughness length of 10^{-4} m over the ocean and 10^{-1} m over land and sea-ice. A parametrization of gravity wave drag (Palmer et al 1986, Slingo and Pearson, 1987) is included.

b. Data

The 5 year control simulation was started from 1979 FGGE data. Fields from the integration to be shown here are meaned over the five three month periods (July, August and September) unless otherwise stated. Verification data from the Meteorological Office operational archive are meaned over 5 austral winters, (July, August and September) 1983 to 1987.

A longer averaging period would be desirable, but the main features of the analyses are similar to longer term climatologies. Karoly and Oort (1987) have compared two recent independent ten year mean data sets of southern hemisphere circulation statistics, one from the Geophysical Fluid Dynamics Laboratory (henceforth referred to as GFDL) and the other prepared by the World Meteorological Centre in

Melbourne (henceforth referred to as ASH). In general, the Meteorological Office analyses are similar to the GFDL and ASH analyses, but in high latitudes there are some systematic differences. Firstly, both the GFDL and ASH analyses are slightly warmer at low levels over Antarctica, and consequently 500 mb heights are slightly greater. This is largely due to the slightly different meaning period (June to August as opposed to July to September), as there is a general reduction in 500 mb heights through the winter. The data of Oort (1983) indicate a reduction of 5 dam within the polar vortex between June and September, whereas Taljaard et al (1969) show a decrease of 15 dam over the same period. Secondly, there is a second maximum in the meridional profile of zonal wind near 55°S which is not present in the GFDL analysis and only hinted at in the ASH data. Karoly and Oort (1987) suggest that the mean flow in the GFDL analyses may be too weak at 55°S, and an earlier climatology (Van Loon et al 1971) shows a weak maximum in zonal geostrophic winds at these latitudes in July, and a more pronounced maximum in October. Bearing in mind these factors, and the different years over which the data sets were collected, it appears that the Meteorological Office data are sufficiently representative of the long term Southern Hemisphere climatology to validate the model.

c. The model antarctic winter climatology

The simulated circumpolar surface pressure trough (Figure 1a) is very similar to that observed (Figure 1b) with two main centres below 980 mb located just off the coast of Antarctica near 90°E and 135°W

respectively. There is a third low pressure centre in the modelled surface pressure field near 0°E as compared to 30°E in the operational analyses. The subtropical high pressure ridge is similar in strength to that observed, except for slightly high values over Australia, so that the general strength of the westerly gradient in mid-latitudes is well represented. 500 mb heights (Figure 2a) are between 6 and 12 dam lower than observed (Figure 2b) over Antarctica. In the operational analyses, there are mid-latitude troughs near 15°E, 100°E and 150°W. In the simulation, there are weak troughs in high-latitudes near 60°E and 150°E, and a more pronounced trough extending into mid-latitudes near 90°W.

In July, simulated surface temperatures over sea-ice (Fig. 3a) are in places 10K or more lower than the observed screen temperatures (Fig. 3b). Note that surface temperatures will tend to be lower than screen temperatures if there is a surface inversion, but this is unlikely to explain all of the discrepancy between Figures 3a and b. Sea-ice thickness is assigned a value of 2 metres at all model sea-ice points. However satellite data on Antarctic sea-ice (Zwally et al, 1979) indicates that a maximum depth is likely to be $1^1/_2$ metres and leads of open water occupy up to 25% of its area. Hence modelled ocean to atmosphere heat fluxes are inhibited and surface heating of the air over sea-ice in the model is likely to be less than observed (Figure 5). Over Antarctica the simulated temperature is very similar to that observed with modelled temperatures in the interior of the continent being -55°C and with a sharp positive gradient over the coastal slopes.

The vertical distribution of zonally averaged temperature from the model (Fig. 6a) is generally close to observed (Fig. 6b) except at upper levels in middle and particularly high latitudes where the model is over 5 K colder than observed. The double jet structure in zonally averaged west to east wind in the Meteorological Office analyses is well reproduced, though as noted above, the double maximum is not present in the GFDL or ASH analyses. The simulated strength is probably excessive above 200 mb.

The strong westerly flow in mid latitudes is evident in the lowest level model winds (σ = 0.987, Fig. 7a). The strength of the mean wind is smallest over sea-ice, due to both the slacker pressure gradient and the larger drag coefficient of the sea-ice. Over Antarctica, the winds are generally weaker than over the open ocean, and are away from the regions of high orography, turning anticyclonically as they reach the edge of the continent. In places, (for example near 85°S 75°W and 80°S 165°W) the winds flow down from the highest regions and follow the contours of the topography, before converging in valleys along the coast. This is similar to the streamline pattern produced by the simple Ball model of katabatic flow based on observed temperature data (Ball, 1960, Parish and Bromwich, 1984) (Fig. 7b). In other areas (for example, near 80°S 150°E the tendency of the winds to follow orographic contours is negligible. The strength of the wind (0-5 m/s) is much less than for observed katabatic winds. This may be because of the thickness of the bottom layer, about 200 m, whereas the strongest katabatic flow occurs near 50 m (Parish, 1984) and the failure of the model orography (Fig. 4) to resolve most of the coastal valleys.

In order to assess the performance of the model on synoptic time scales (fluctuations with periods up to about 6 days) we have calculated high pass filtered variances of 500 mb height calculated from the difference between the variance of daily fields and the variance of 3 day mean fields. The variances were calculated separately for each year and then averaged over the 5 years of the simulation (Figure 8a) and 5 years of the operational analyses (Figure 8b). The intensity and position of the circumpolar storm track is simulated well, except perhaps from about 135°W to South America where the simulated maximum is displaced equatorward.

We have shown that the model produces a realistic simulation of the antarctic winter. A reported deficiency of many models, for example those at GLAS (Goddard Laboratory for Atmospheric Sciences, Herman and Johnson 1974), NCAR (National Centre for Atmospheric Research, Schlesinger 1984), CCC (Canadian Climate Centre, Boer et al 1984)) and the University of Melbourne (Simmonds 1983) and the British Meteorological Office 5-layer model (MH) is the poor simulation of the antarctic circumpolar trough in the Southern Hemisphere winter. Simulated pressures are 5 to 15 mb higher than observed. The trough is generally positioned equatorwards of its observed position although the strength and position of the maritime highs are well simulated. Consequently the strong observed gradient between 45 and 60°S is not evident and surface westerly winds are weaker than observed. The surface temperature over the antarctic plateau is overestimated by most models, the GLAS model producing temperatures 30° C higher than observed. This is thought to be due to a poor surface energy

balance. Pressures over the continent itself are overestimated by 20 to 50 mb, although this may be due to errors in the method of reduction to mean sea level.

The present model shows a substantial improvement in these areas, especially in the simulation of low level winds over the antarctic continent. Some errors are still evident, particularly upper tropospheric temperatures in high latitudes being colder than observed (the 'cold pole' problem) giving an overly strong westerly jet. Surface temperatures over sea-ice are also still too cold. Nevertheless, the model's climatology is closer to observed than those used in previous antarctic sea-ice anomaly experiments.

3. The sea-ice anomaly experiment and comparison with MH

In the previous section we established that the model produces a realistic simulation of present day climate in high southern latitudes. This is necessary if we are to have confidence in the response of the model to climate perturbations. In this section we discuss results from an anomaly experiment in which all Southern Hemisphere sea-ice equatorward of 67.5°S was removed (Figure 4) and replaced by ocean at 271.2 K. This is similar to the experiment in MH, except that here the sea-ice extents in the control simulation are less extensive (and more realistic) so that the perturbation is smaller. As in MH, three anomaly simulations were carried out, each run for 4 months commencing on the 1st June of years 1, 2 and 3

of the control simulation respectively. The results are averaged over the final three months of the three integrations, and the experiment is called ICEANOM.

The immediate effect of removing sea-ice is to raise the surface temperature (Fig. 9a). The changes are less widespread than in MH, reflecting the smaller reduction in ice area, and the magnitude of the maximum changes is smaller, as a result of lower temperatures over the ice sheet in MH's control simulation. The increase in surface temperature produces an increase in turbulent fluxes of sensible heat in the region where open water is exposed (Fig 9b) and decreases to the north as explained in MH. The increase in the flux of sensible heat is much smaller than found by MH, probably because the rise in surface temperature is smaller, although differences in boundary layer formulation of the two models could contribute. On the other hand, the changes in the flux of latent heat are similar in both experiments (Fig. 9c). This may be explained as follows. Very low temperatures are found over sea-ice in both control simulations; this leads to negligible saturation vapour pressures at the surface and at low levels, and hence a negligible rate of evaporation. When sea-ice is removed the saturation vapour pressure is calculated at a temperature of 271.2K in each anomaly experiment. As the dominant term in the evaporation rate is the surface saturation vapour pressure, in both cases the evaporation rate rises from near zero to approximately the same value.

The zonally averaged atmospheric warming is largely confined to below 700 mb and within 10° of latitude of 65°S (Fig. 10a). In MH, the warming in the lowest model layer spreads across Antarctica where it exceeds 7K, whereas here one may speculate that the strong katabatic winds (Fig. 7a) restrict the penetration of warm low level air across the antarctic plateaux. Above the boundary layer the presence of a warm atmospheric anomaly near 65°S gives rise to an easterly wind anomaly to the north, and a westerly anomaly to the south, as expected from the thermal wind relationship (Figure 10b). Near the surface, however, the westerly wind anomaly extends north of 65°S, consistent with the changes in surface pressure (Fig. 11, solid line).

The maximum fall in zonally averaged surface pressure coincides with the latitude of the greatest increase in surface and atmospheric heating, that is, over the ocean in the row next to the new ice edge. The maximum fall is smaller and less consistent from month to month than in MH, probably because of the smaller change in atmospheric heating. The main falls in surface pressure are found near the new ice edge near 30°E (Fig. 12) just downstream from the largest decrease in ice extent (see Fig. 4). MH found the largest falls near 60°E, consistent with the largest changes in ice extent in their experiment being situated further east. In the present experiment, there are further falls in surface pressure near 135°E (135°W in MH) and rises in lower latitudes east of South America and near 45°W (Fig. 12) which are significant at the 90% level of confidence or above. MH found a significant rise in pressure over Antarctica whereas here there are decreases, although they are not statistically significant. In general the longitudinal distribution of the changes in the two

experiments is quite different, but in each case there is a well defined fall in pressure in the region from which ice was removed. This confirms that the difference between the surface pressure response reported by MH and by S and SD is not an artifact of the relatively poor control simulation in MH.

4. Mechanical and Thermal forcing

As noted in the introduction, the Meteorological Office models use a larger roughness length over sea-ice than over the ocean, so that replacing sea-ice by ocean is accompanied by a substantial reduction in momentum drag coefficient (Figure 13). The effect of this change is isolated in an additional experiment described in this section.

Three further anomaly integrations (Experiment SMOOTH) were run using the same climatological sea-ice extents as in the control, but with the surface roughness over all Southern Hemisphere sea-ice equatorward of 67.5°S reduced from 10^{-1}m to 10^{-4} m. The reduction in surface roughness leads to a reduction in the surface exchange coefficients for momentum and sensible and latent heat (Mitchell, Wilson and Price, 1985). For neutral conditions, this corresponds to a change from 7.5×10^{-3} to 1.2×10^{-3} in the momentum exchange coefficient at 10 m. The zonally averaged changes in surface temperature and the fluxes of sensible and latent heat are small compared to those in the sea-ice anomaly experiment (ICEANOM) described in the previous section (Fig. 9). However, the reduction in the model surface drag coefficient in the region of the anomaly in SMOOTH is very similar to that found in ICEANOM (Fig. 13). A reduction in surface drag (in the

absence of changes in surface pressure gradient) will reduce the low level convergence from the circumpolar westerlies into high latitudes. The surface pressure in SMOOTH is reduced in high latitudes whereas equatorward of the anomaly pressure is increased (Fig. 11, dashed line). This leads to a slight increase (up to 1 m/s) in westerly flow and a small cooling (up to 1 K) in the boundary layer near 65°S (not shown). The falls in surface pressure over parts of the Weddell and Ross Seas, to the west of South America and near 135°E (Fig. 14) are significant at the 90% level of confidence. In summary, the main contribution of reducing surface roughness is a reduction in surface pressure of 2 mb or so in high latitudes.

As noted above, the changes in surface drag coefficient in SMOOTH and ICEANOM are very similar (Fig. 13), whereas the changes in the turbulent heat fluxes in SMOOTH, unlike those in ICEANOM, are negligible (Fig. 8). This indicates that the differences between ICEANOM and SMOOTH are mainly due to the increased surface heating in ICEANOM and should therefore be similar to those found in S and SD. The zonally averaged temperature changes (not shown) are very similar to those in Fig. 10a, but the changes in zonal wind around the latitudes of the ice anomaly (Fig. 15) are almost equivalent barotropic as in S and SD. As in MH, the reductions in surface pressure occur in the main region from which sea-ice is removed (Fig. 16) but are only significant at the 90% level of confidence over a limited region near the Greenwich meridian. The geographical distribution of the changes differs from that in S who found increases in surface pressure around Antarctica extending from 45°E eastwards to the Drake Passage, and from SD who reported significant increases in pressure where sea-ice had

been removed near 30°E, and further increases west of the Drake Passage. These discrepancies are undoubtedly due in part to both the different sea-ice anomalies used within the three experiments, and to the inherent variability of the simulated circulation at these latitudes.

The analysis above indicates that both reducing the surface roughness (as in SMOOTH compared with the control) and increasing the turbulent heating from the surface (as in ICEANOM compared with SMOOTH) produce local reductions in surface pressure. Combining these effects produces larger and more significant effects in the region from which sea-ice has been removed (compare Figs. 14 and 16 with Fig. 12) though the increase in sample size and hence the number of degrees of freedom also contributes. S and SD do not include a change in surface roughness in their experiments, and this appears to explain much of the discrepancy between their results and those reported here and by MH.

5. Summary and concluding remarks

Because of the availability of daily global analyses, we have been able to assess both the mean state and the synoptic variability during the antarctic winter as simulated by a recent version of the Meteorological Office 11-layer model. The low level pressure patterns and winds are in excellent agreement with observations. The model is slightly too cold over Antarctica and at the surface over sea-ice. The latter error may be due to the assumption of a 2 m ice thickness in estimating the surface temperature, where in reality the ice is thinner and there are numerous leads. At 500 mb, the simulated vortex has two broad troughs, whereas

three are present in the analyses. The position and intensity of the storm tracks, diagnosed from the variance of 500 mb height derived using a crude high pass filter is very similar to that observed.

In the sea-ice anomaly experiments that have been considered, there is a reduction in surface pressure over the largest anomalies, and in the Meteorological Office experiments the reduction is most pronounced on the seaward side of the new ice limits. In the present experiment, the reduction in surface roughness due to replacing sea-ice with open ocean contributed to a general reduction in surface pressure in high latitudes. The increase in turbulent heating over the sea-ice anomaly produced a more local reduction in surface pressure. These two components of the changes in surface pressure were only marginally statistically significant when considered individually, but the combined effect was highly significant although this increase in significance is partly to do with sample size. In the experiments of Simmonds (1981) and Simmonds and Dix (1986), only the turbulent heat fluxes were altered, so it is not surprising that they found smaller and less significant changes in surface pressure.

Most atmospheric models assume a compact 2 m ice cover for both the Arctic and antarctic. In reality the antarctic sea-ice is generally thinner than 2 m and contains a substantial fraction of open water. As a result, it is likely that the insulating effect of antarctic sea-ice is exaggerated in current models, and the 'thermal forcing' in the antarctic sea-ice anomaly experiments is much greater than would occur in practice. On the other hand, there is considerable evidence that the surface

roughness over the seasonal sea-ice in the southern hemisphere is much greater than over the open ocean. Hence this should be represented in models used to determine the effect of sea-ice anomalies, whether for seasonal forecasting or assessments of longer term climate change. Note that the values used over sea-ice here are larger than observed.

Finally, the large geographical variations between the responses of the different models considered to a relatively simple perturbation should concern those attempting to use results from numerical simulations to assess regional impacts of climate change. There are several factors which may contribute to these variations.

1. The inherent variability of the simulated climate in high latitudes. This can be overcome by increasing the length of the control and anomaly simulations.

2. The longitudinal distribution of the atmospheric response may be sensitive to the size and position of the anomaly. This is known to be true for other perturbations, such as changes in tropical sea surface temperatures associated with El Nino.

3. The simulated response is dependent on the control simulation and the model formulation. This was discussed in the introduction, and further illustrated here by the effect of using different formulations of surface roughness over sea-ice. It is vital that the mechanisms producing simulated changes in climate are well understood and authenticated.

These are all factors which must be taken into account by those comparing results from different models, or using results from an individual model.

6. Acknowledgments

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Figure Captions

1. Pressure at mean sea level, averaged for July, August and September. Contours every 4 mb.

   (a) Simulated

   (b) Meteorological office operational analyses (1983-7).

2. Height of 500 mb geopotential surface, averaged over July, August and September. Contours every 10 dam.

   (a) Simulated

   (b) Meteorological Office operational analyses (1983-7).

3. (a) Simulated surface temperature (July to September). Contours every 5°C.

   (b) Climatological surface air temperature (July). From Schutz and Gates, 1972. Contours every 5°C.

4. Model topography. (Contours every 500 m), and sea-ice extents for August. The inner stippled area denotes the sea-ice extents used in experiment ICEANOM.

5. (a) Zonally averaged surface fluxes of sensible heat (WM^{-2}). Solid line, simulated; dashed line, from climatological data (Schutz and Gates, 1972).

(b) As a, but for latent heat.

6. Zonally averaged temperatures (solid lines, every 5°C) and winds (dashed lines, every 5 m/s) for July to September. The vertical coordinate is pressure.

(a) Simulated

(b) Meteorological Office operational analyses (1983-7).

7. Low level winds.

(a) Model winds at $\sigma = 0.987$, July to September. The wind speed is proportional to the length of the arrows, and the scale is given in the bottom left hand corner.

(b) Time averaged near surface wintertime streamlines of cold air drainage over Antarctica, derived from the Ball model (Parish and Bromwich, 1987. Reprinted by permission from Nature Vol. 328 pp. 51-54. Copyright (C) 1987 Macmillan Magazines Ltd). The thin lines are ice sheet elevations at an interval of 100 m.

8. High pass filtered variance of 500 mb geopotential height contours every dam), averaged over the period July to September.

(a) Simulated

(b) Meteorological Office operational analyses (1983-7).

9. Simulated changes in zonally averaged quantities in experiments ICEANOM (solid line) and SMOOTH (dashed line), averaged over the period July to September.

(a) Surface temperature (K)

(b) Sensible heat (Wm^{-2})

(c) Latent heat (Wm^{-2}).

10. Zonally averaged changes due to reducing sea-ice extents (ICEANOM).

(a) Temperature. Contours at 1K, 2.5 K and then every 2.5 K. Areas of decrease are stippled.

(b) Zonal component of wind. Contours at every 1 m/s, and areas of decreased westward flow are stippled.

11. Simulated changes in zonally averaged surface pressure (mb).

Solid line, ICEANOM; dashed line, SMOOTH.

12. Changes in surface pressure due to reducing sea-ice extents (ICEANOM-control). Contours are at every 2 mb, areas of decrease are stippled, and changes locally significant at the 10% level of confidence are shaded heavily.

13. Zonally averaged surface exchange coefficients for momentum. Solid line, control simulation; dashed line, ICEANOM; chain dotted line, SMOOTH.

14. As 12, but for SMOOTH - control.

15. As 10b, but for ICEANOM - SMOOTH.

16. As 12, but for ICEANOM - SMOOTH.

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CONTROL
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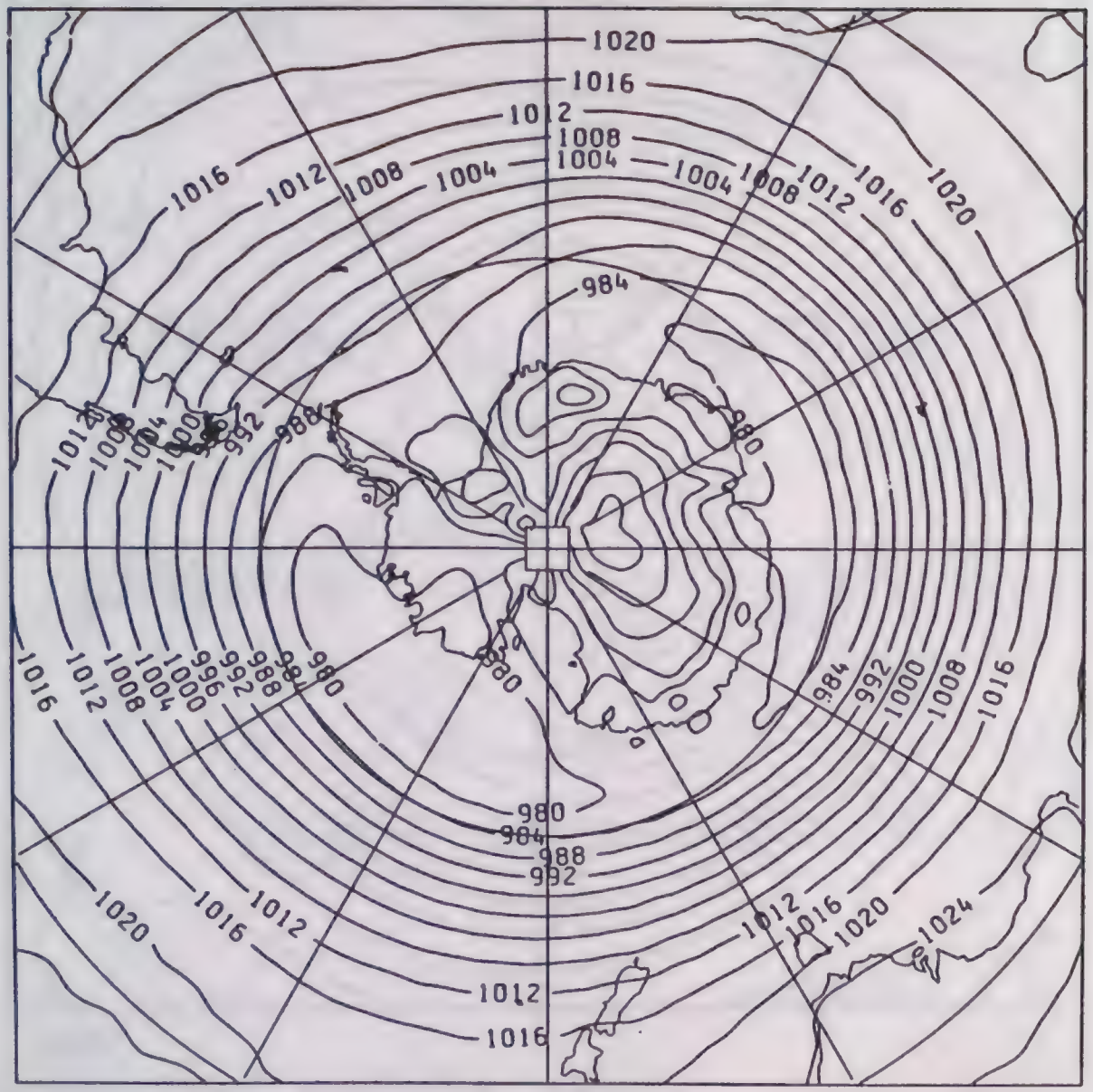


FIGURE 1(a)

OPERATIONAL ANALYSES 1983-87

PMSL

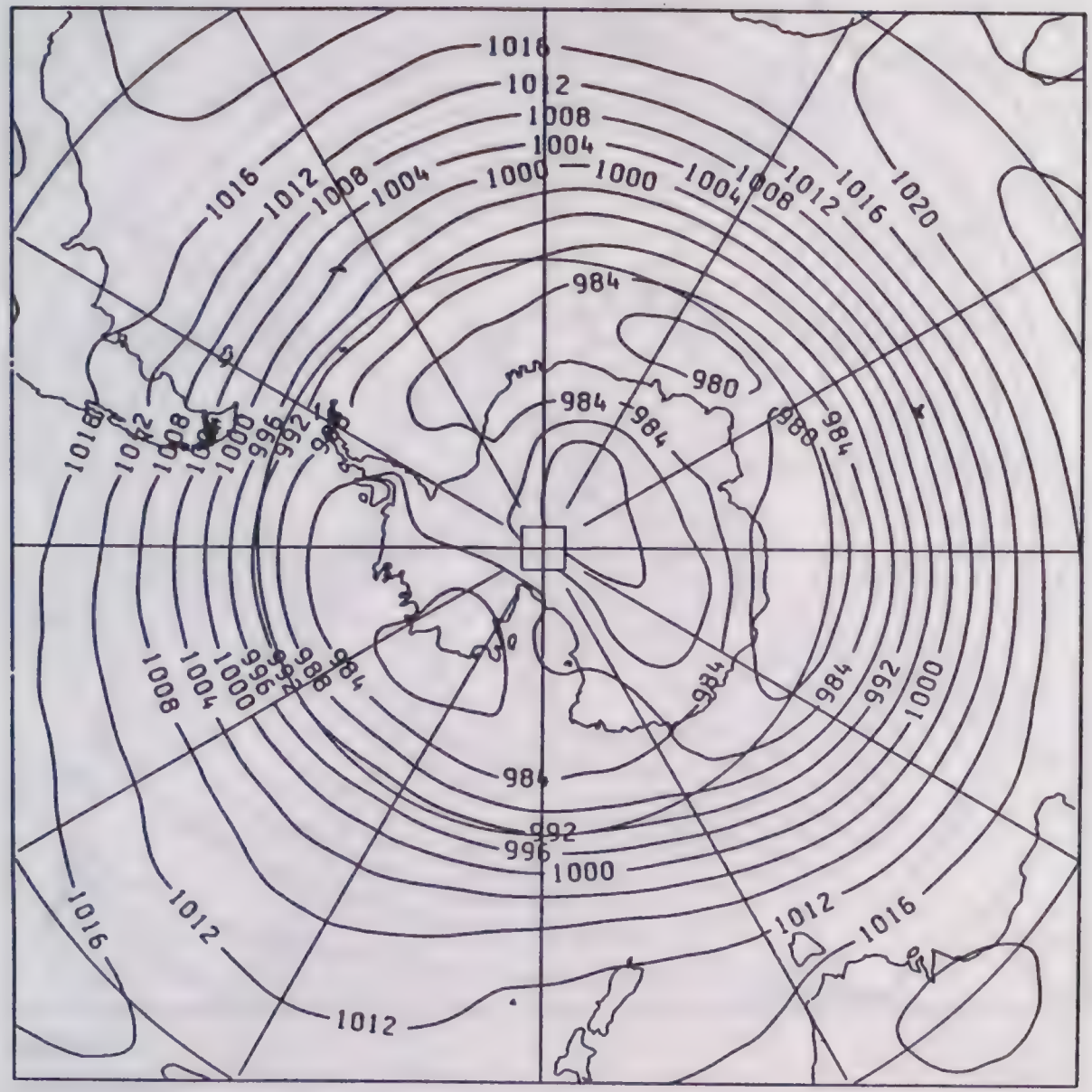


FIGURE 1(b)

CONTROL
500MB HT

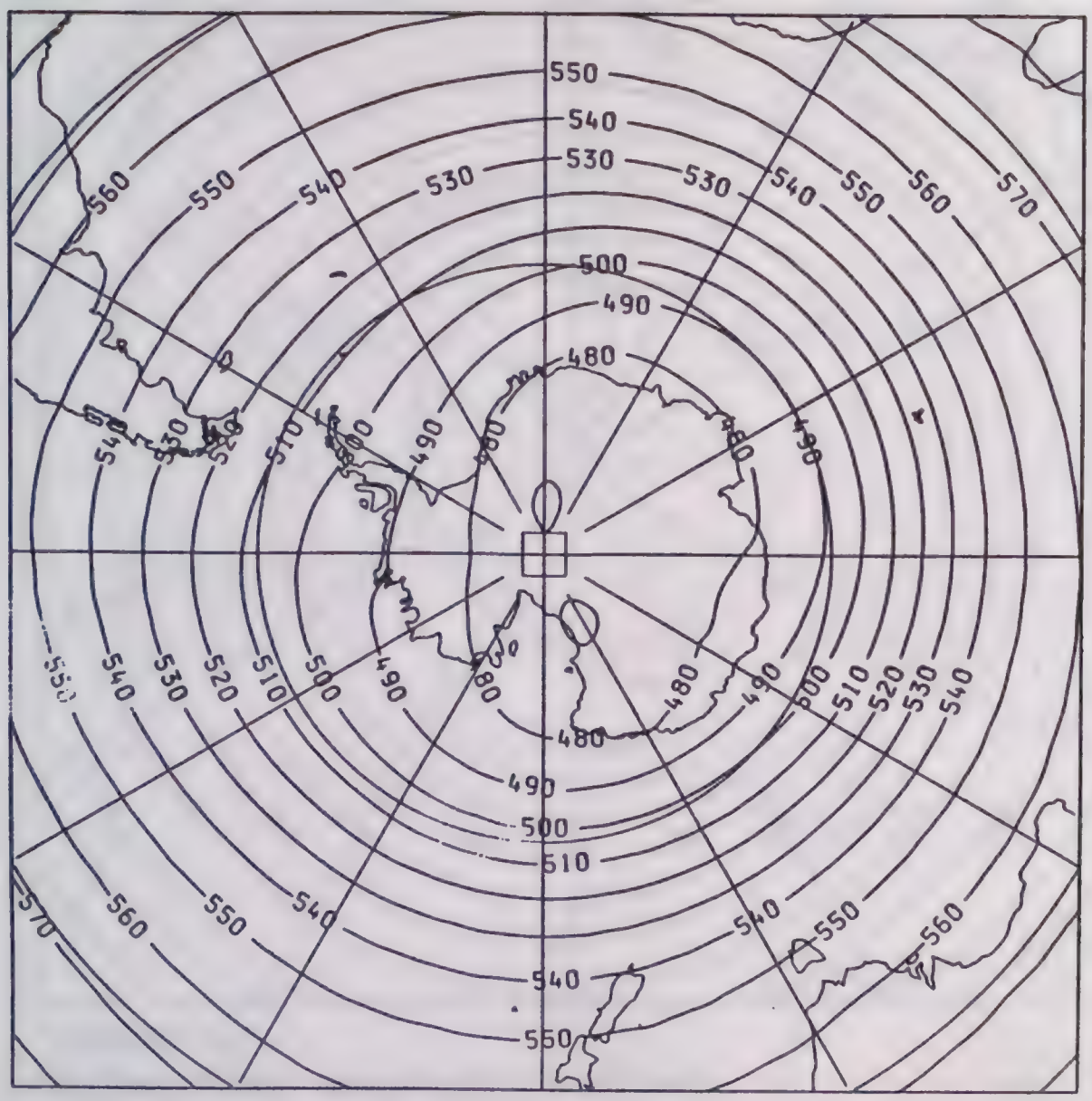


FIGURE 2(a)

OPERATIONAL ANALYSES 1983-87
500MB HT

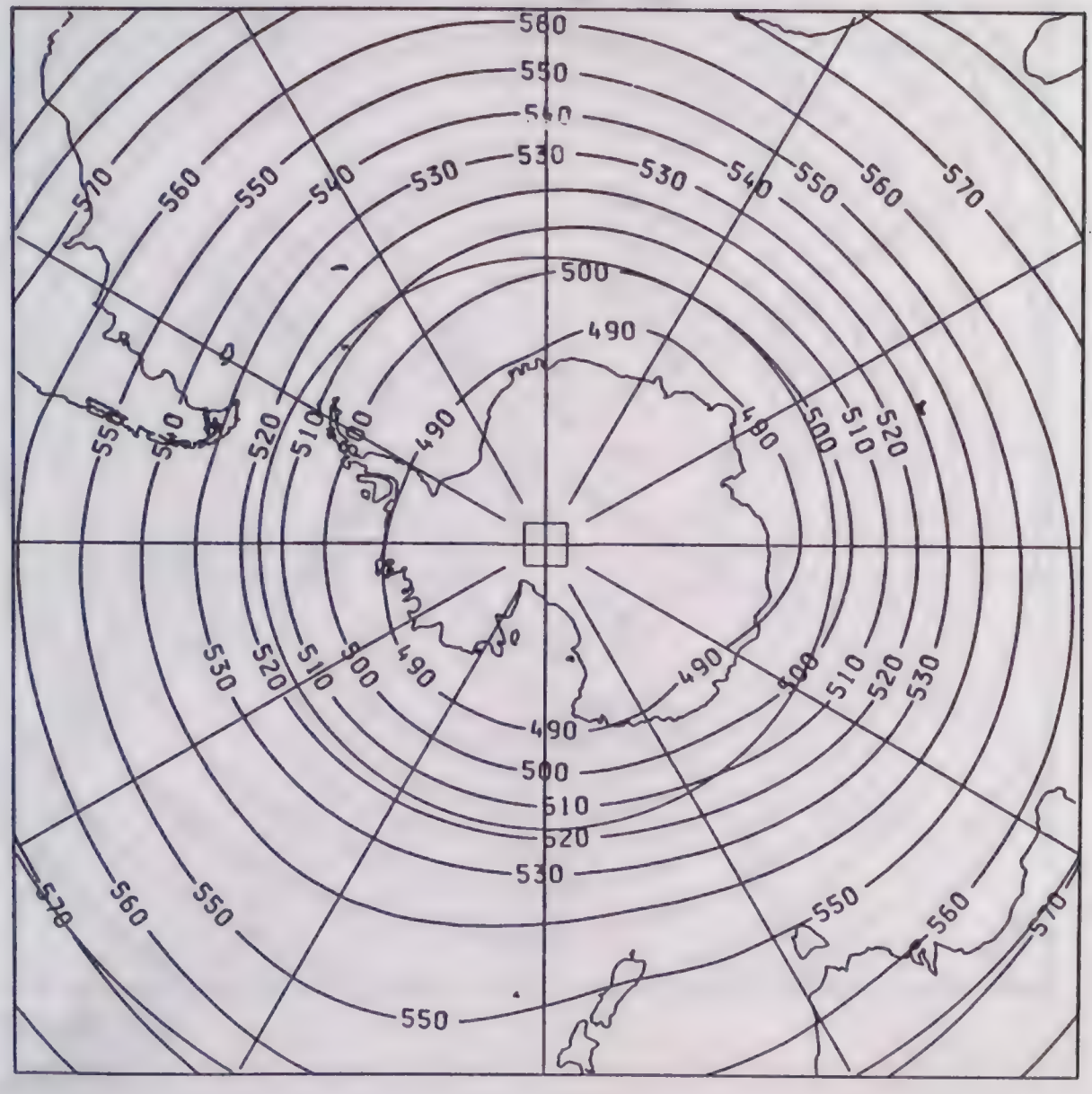


FIGURE 2(b)

CONTROL
SURFACE TEMPERATURE

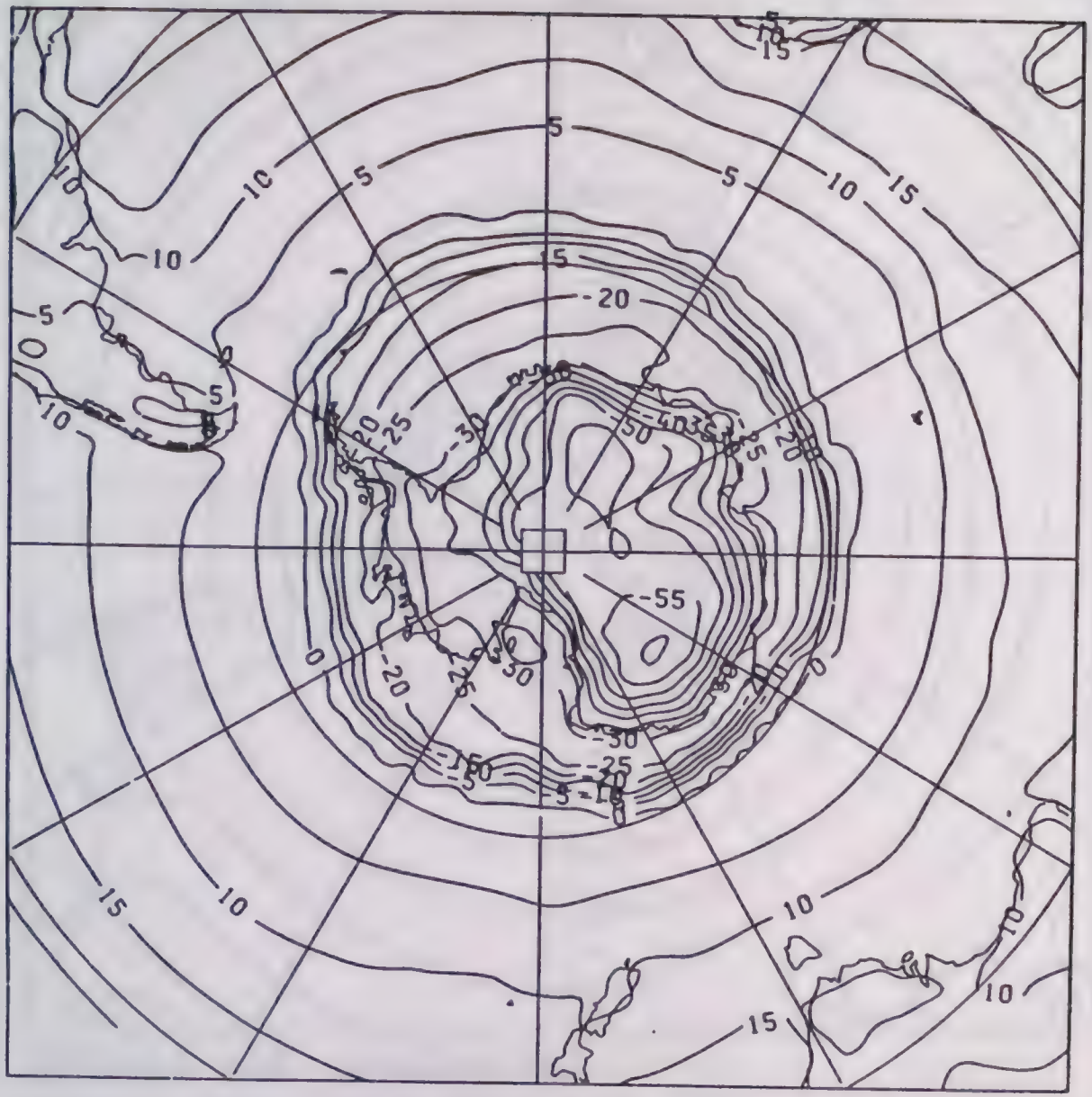


FIGURE 3(a)

SCHUTZ AND GATES DATA FOR JULY
SURFACE AIR TEMPERATURE

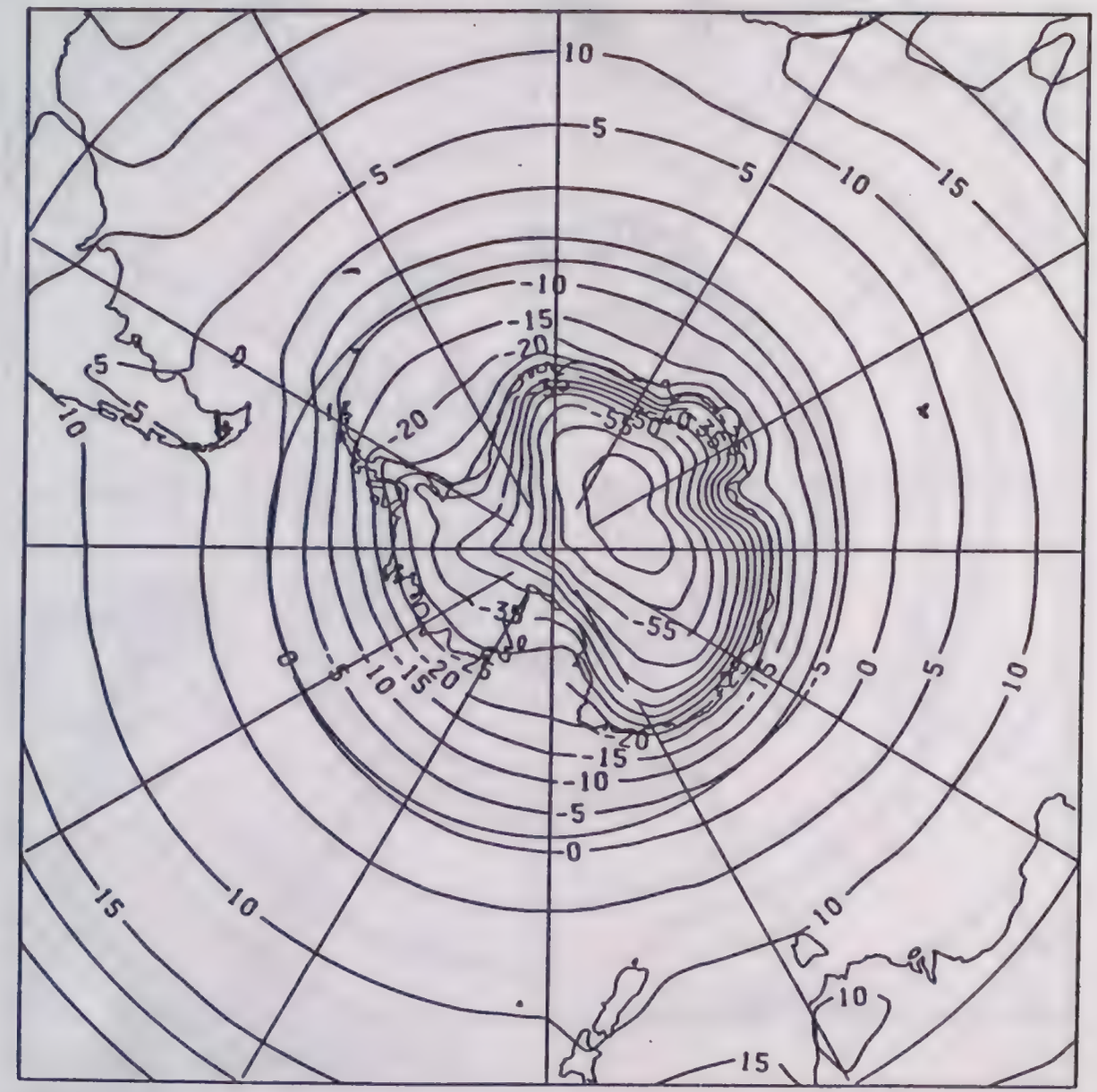
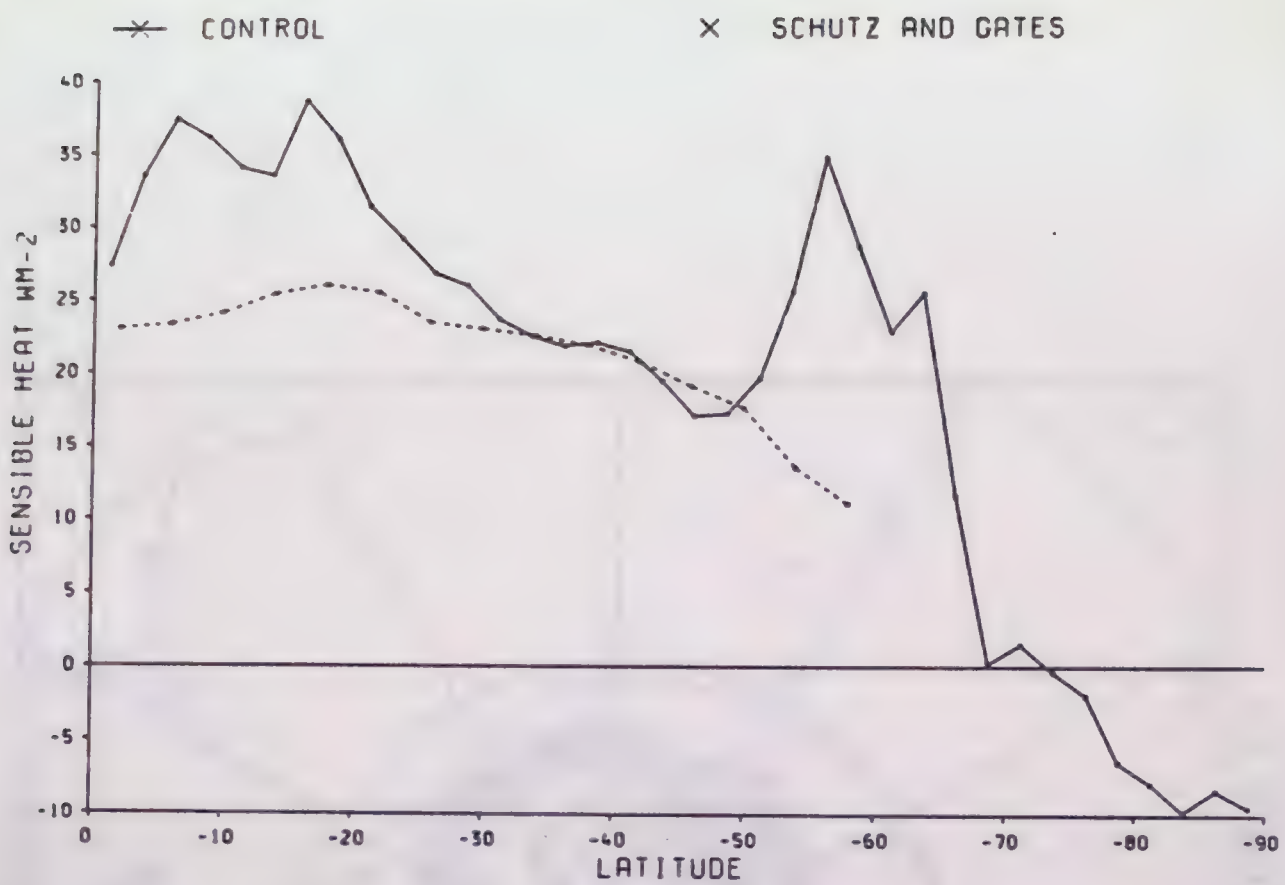


FIGURE 3(b)

OROGRAPHY AND JAS MEAN SEA-ICE EXTENTS
ICEANOM SEA ICE EXTENTS STIPPLED

FIGURE 4

(a) ZONALLY AVERAGED SENSIBLE HEAT



(b) ZONALLY AVERAGED LATENT HEAT

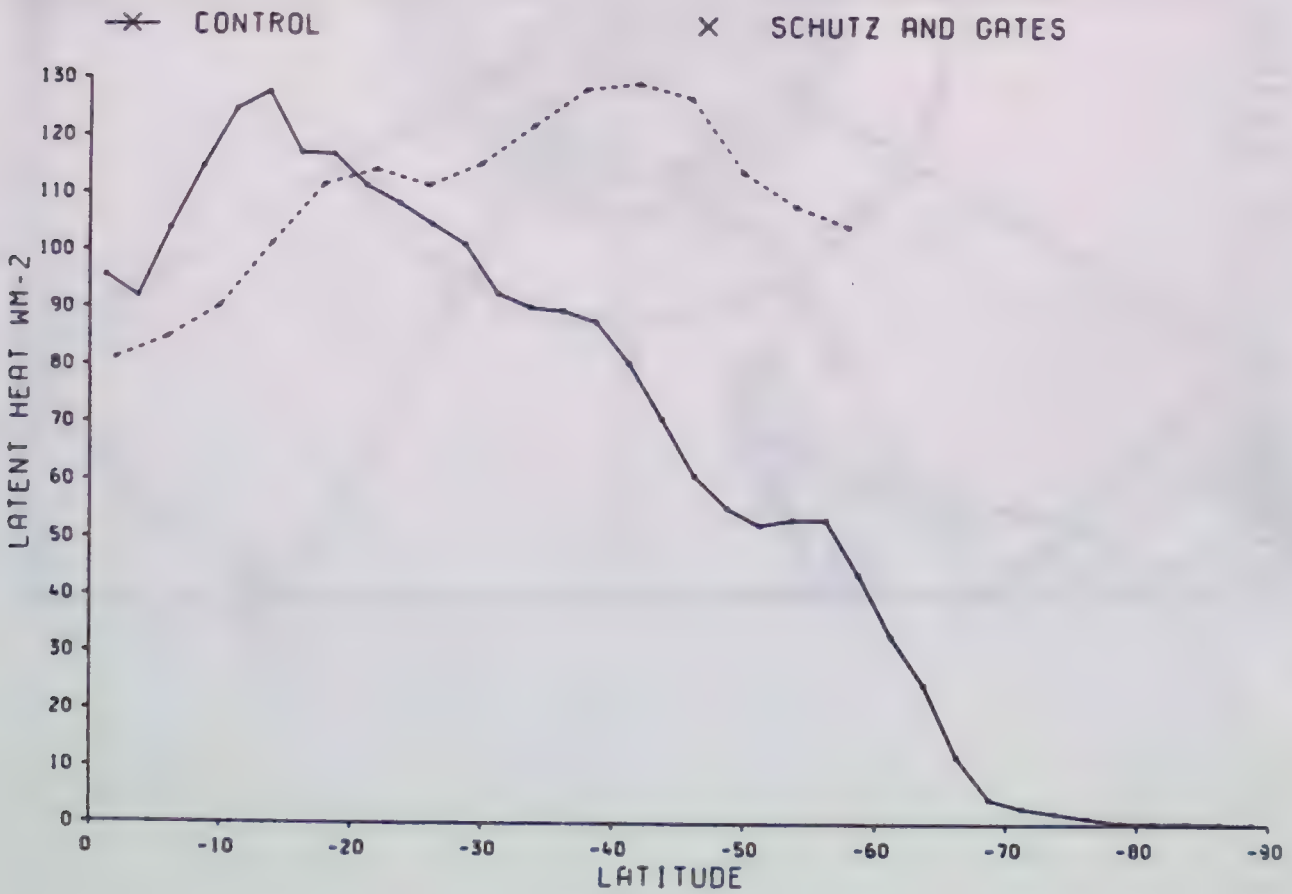
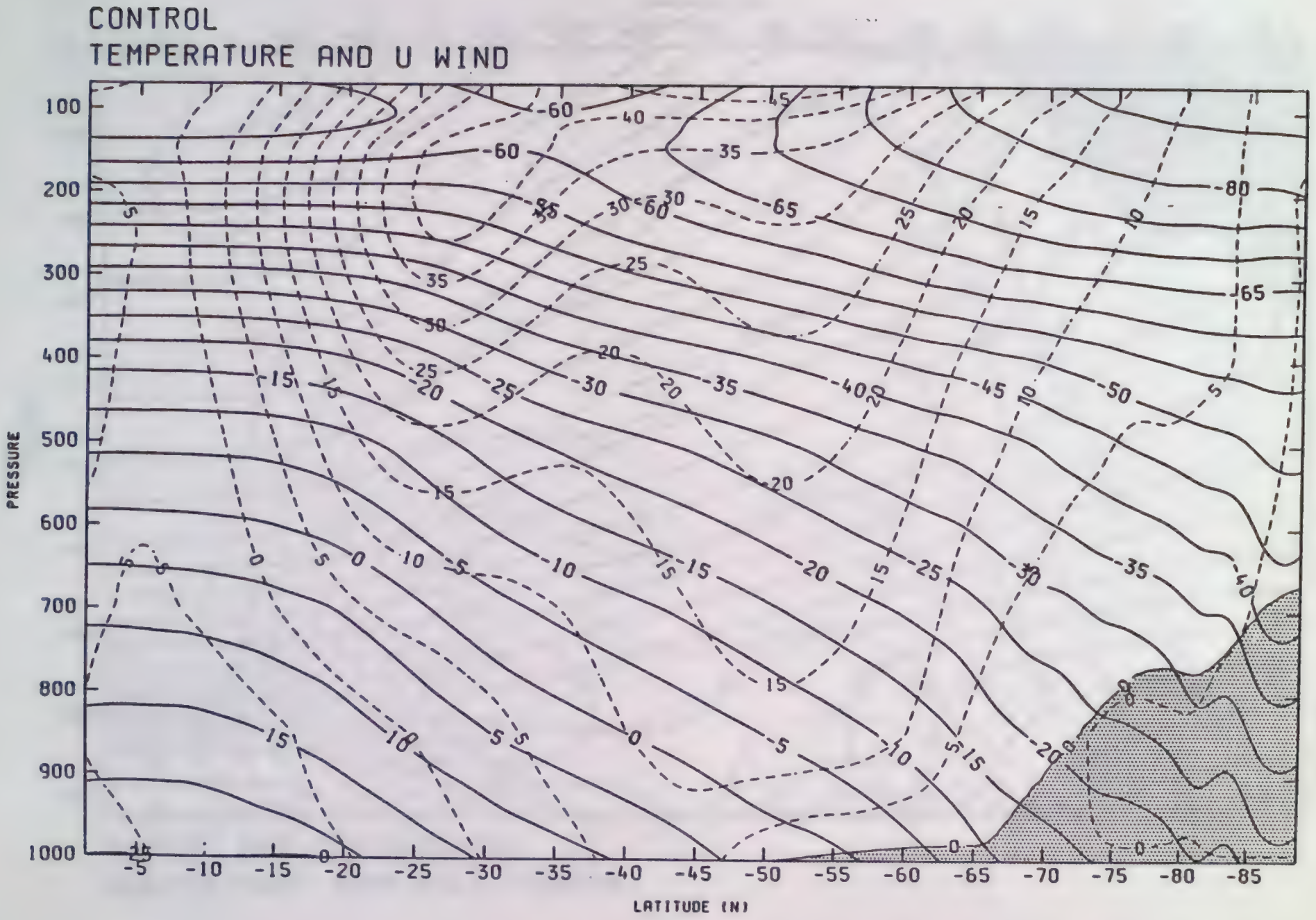


FIGURE 5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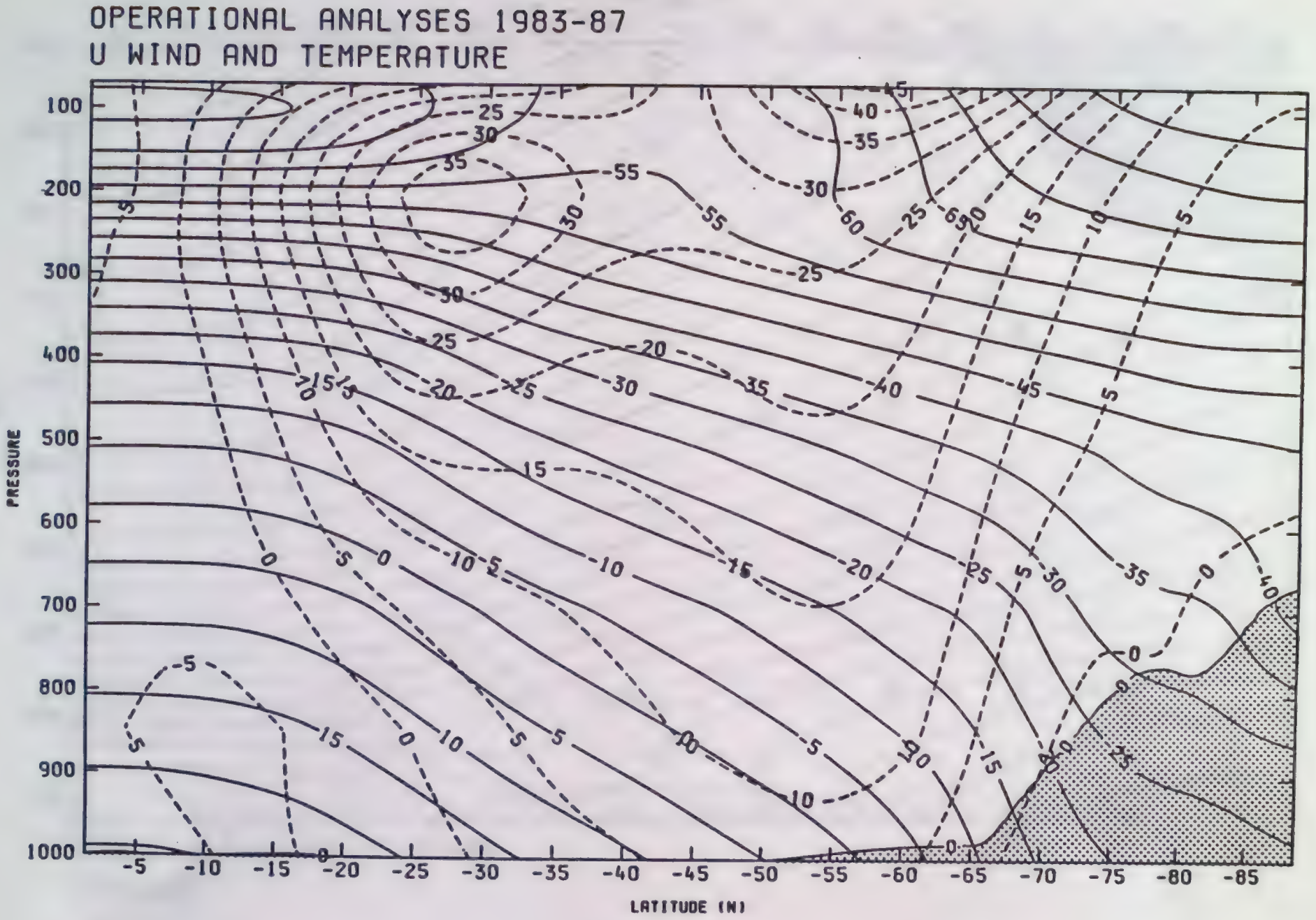
LATITUDE (N)

FIGURE 6(a)


OPERATIONAL ANALYSES 1983-87
U WIND AND TEMPERATURE
PRESSURE
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700
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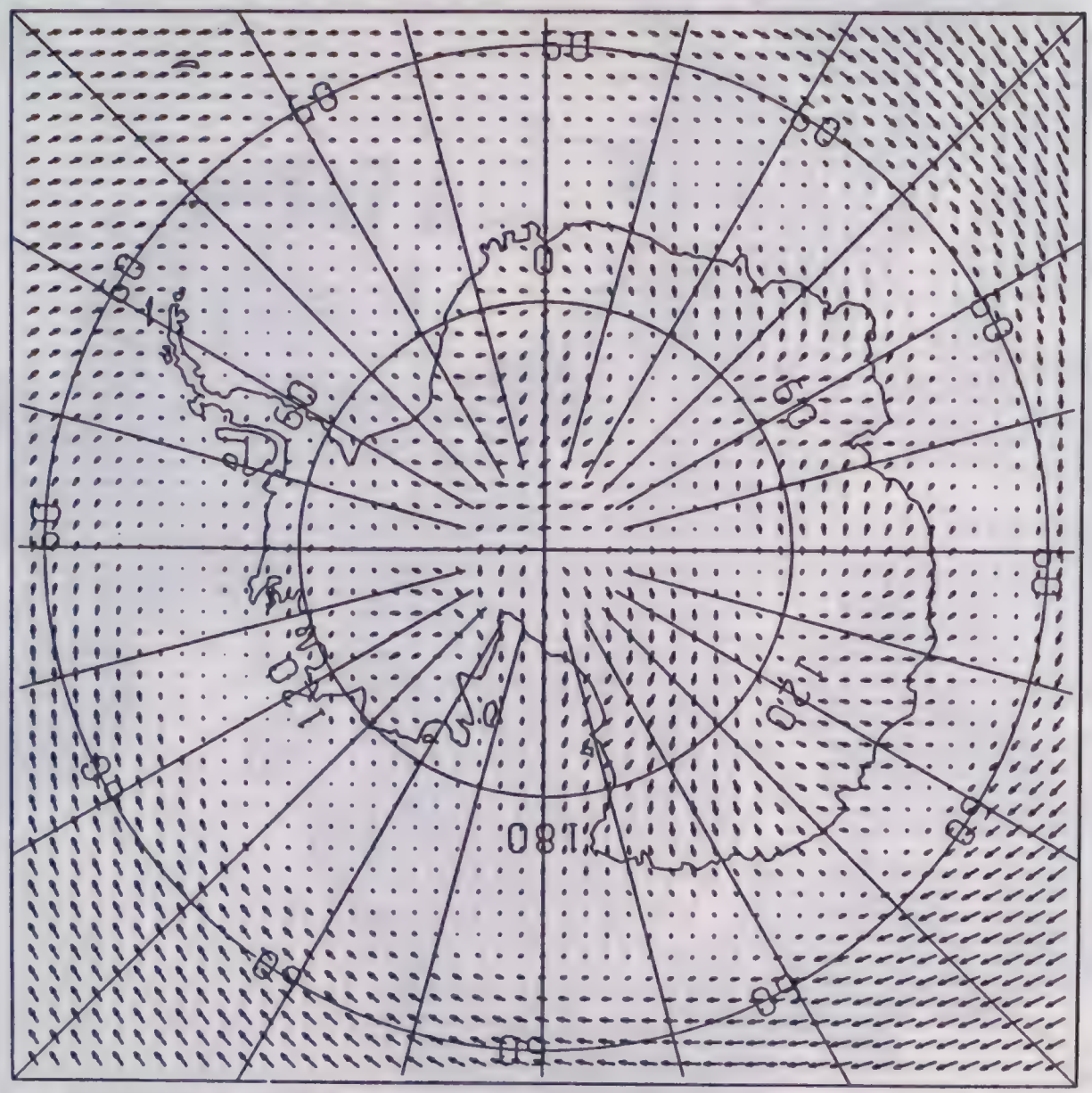
-5 -10 -15 -20 -25 -30 -35 -40 -45 -50 -55 -60 -65 -70 -75 -80 -85
LATITUDE (N)

FIGURE 6(b)

CONTROL

LOWEST LEVEL WIND VECTORS



— REPRESENTS 20 M/S

FIGURE 7(a)

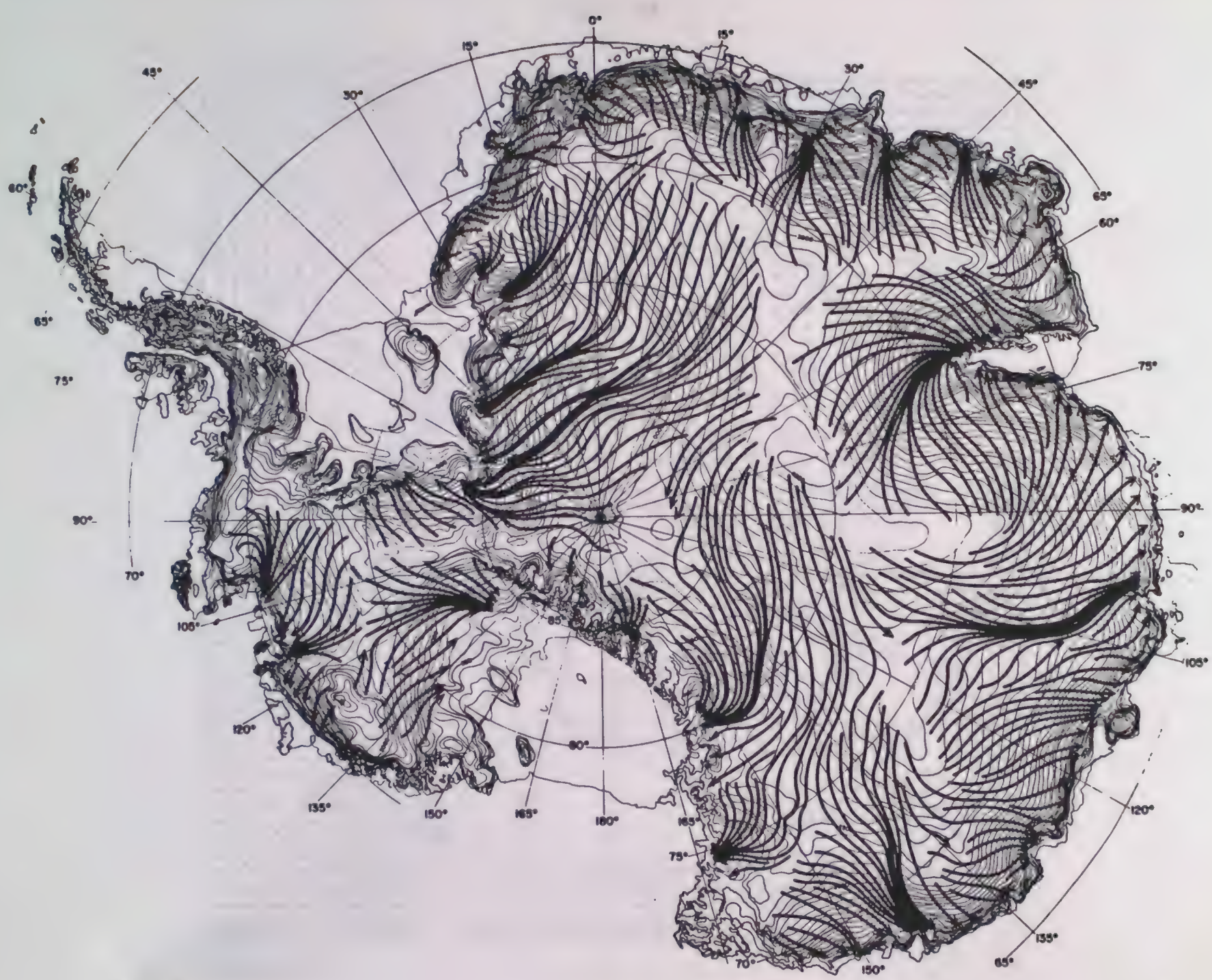


FIGURE 7(b)

CONTROL 5YR JAS MEAN
HIGH PASS FILTERED RMS 500MB HEIGHT

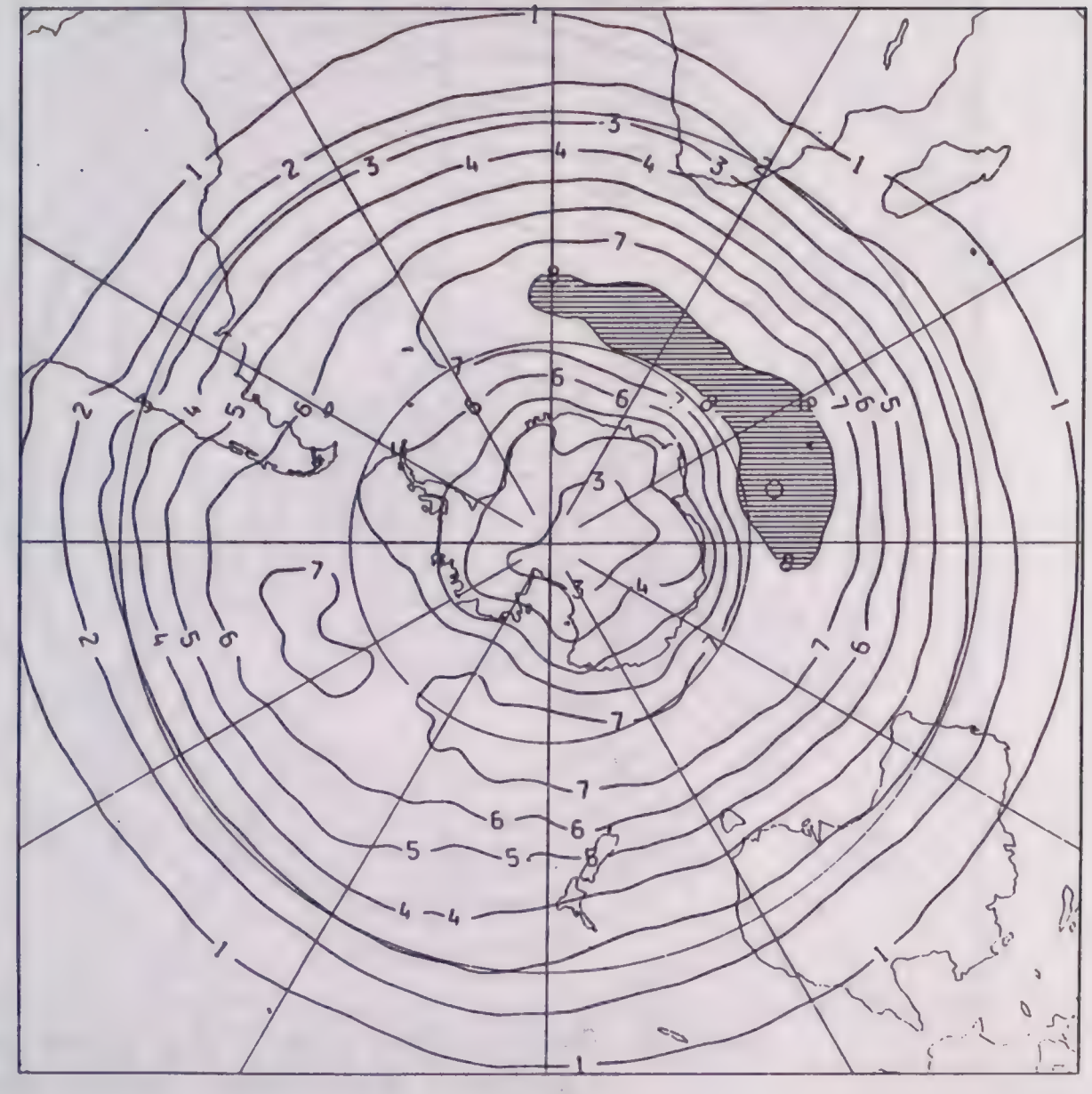


FIGURE 8(a)

OPERATIONAL ANALYSES 1983-87
HIGH PASS FILTERED RMS 500MB HEIGHT

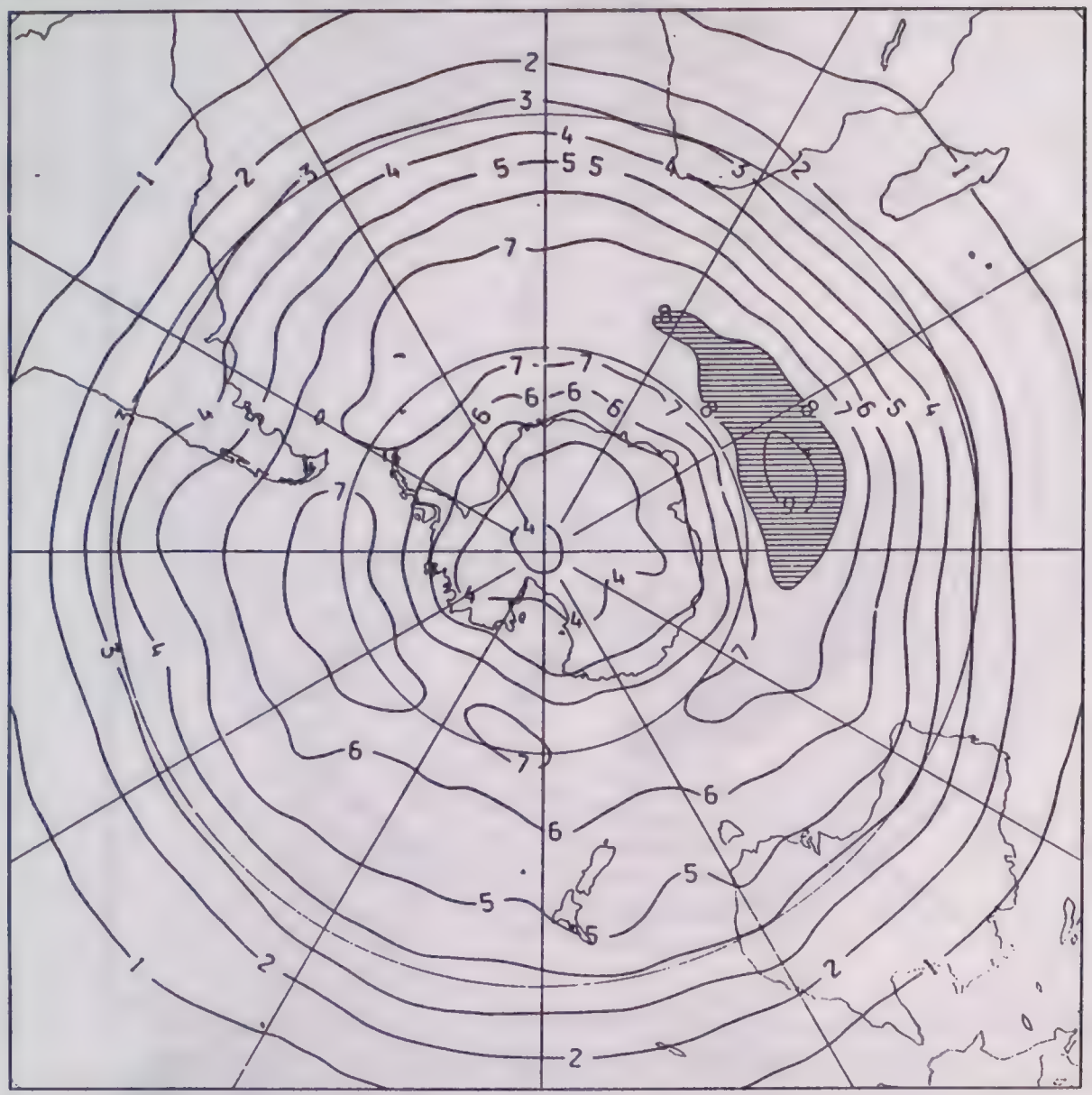
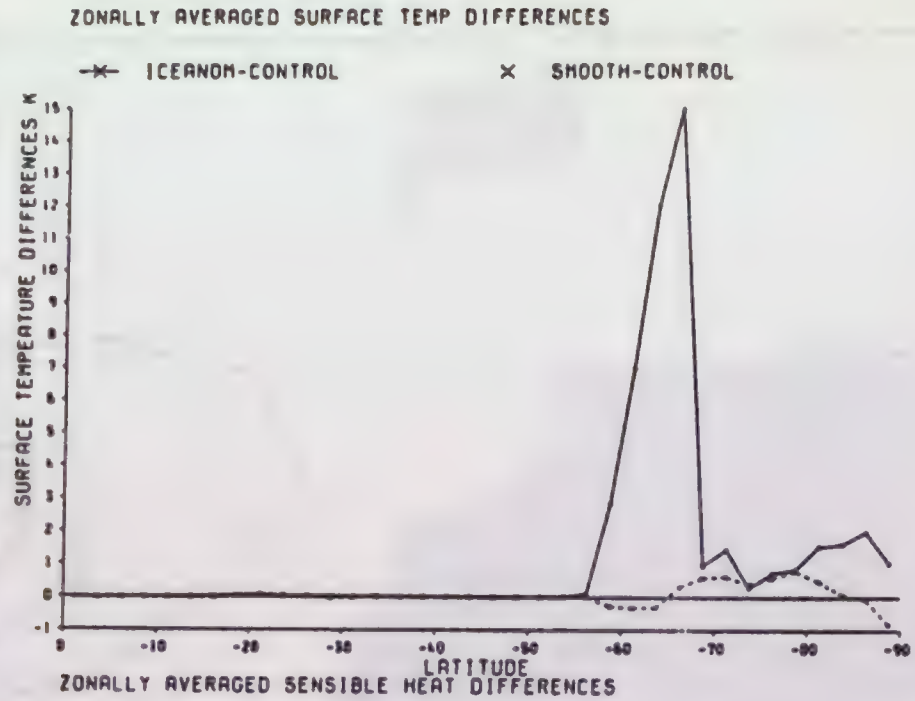
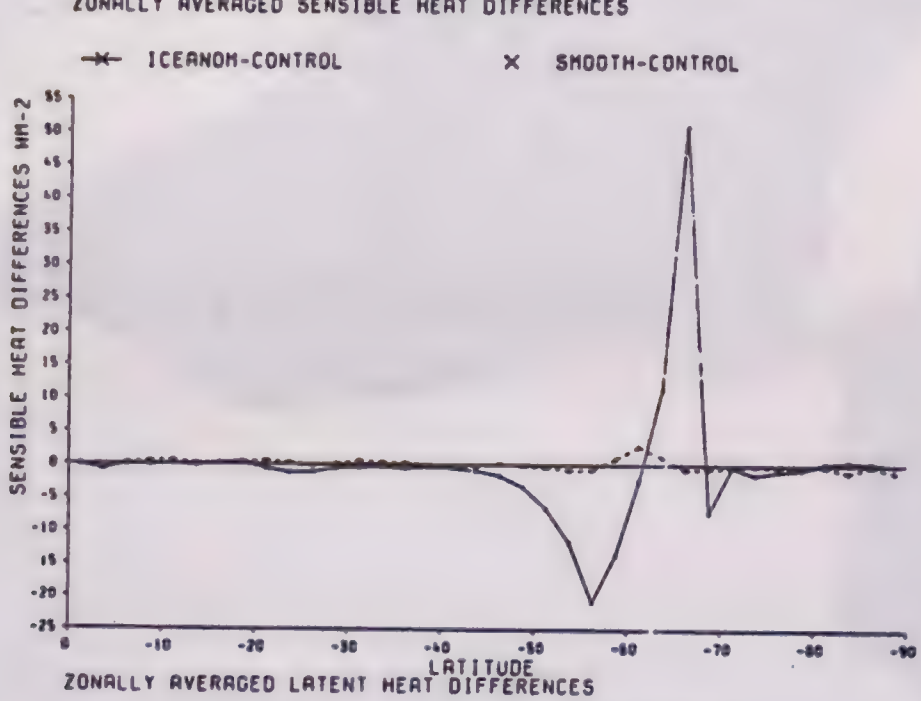


FIGURE 8(b)

(a)


ZONALLY AVERAGED SURFACE TEMP DIFFERENCES
ICEANOM-CONTROL
SMOOTH-CONTROL
SURFACE TEMPERATURE DIFFERENCES K
LATITUDE

(b)


ZONALLY AVERAGED SENSIBLE HEAT DIFFERENCES
ICEANOM-CONTROL
SMOOTH-CONTROL
SENSIBLE HEAT DIFFERENCES WM-2
LATITUDE

(c)

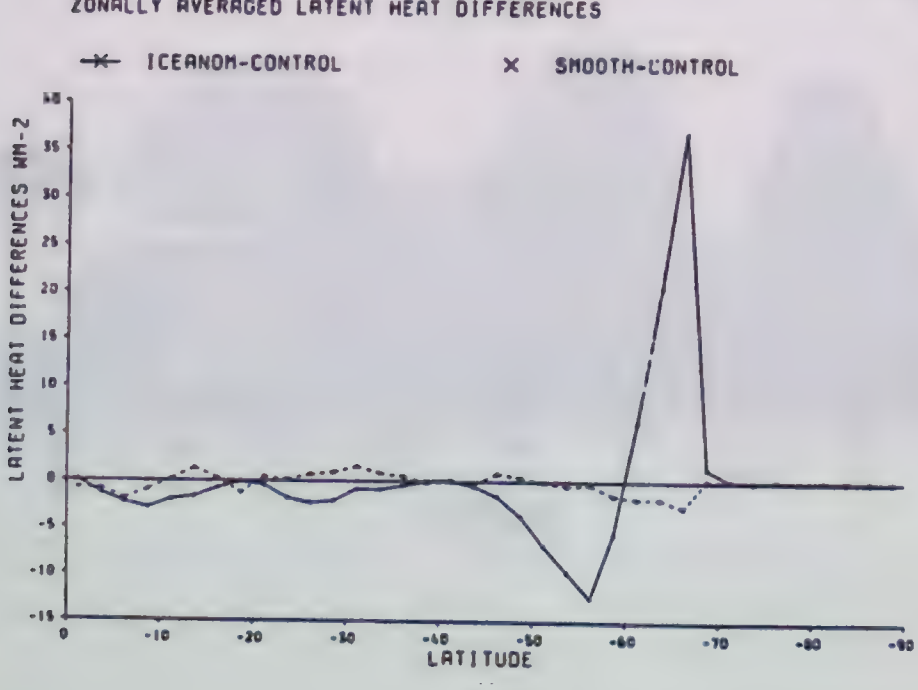
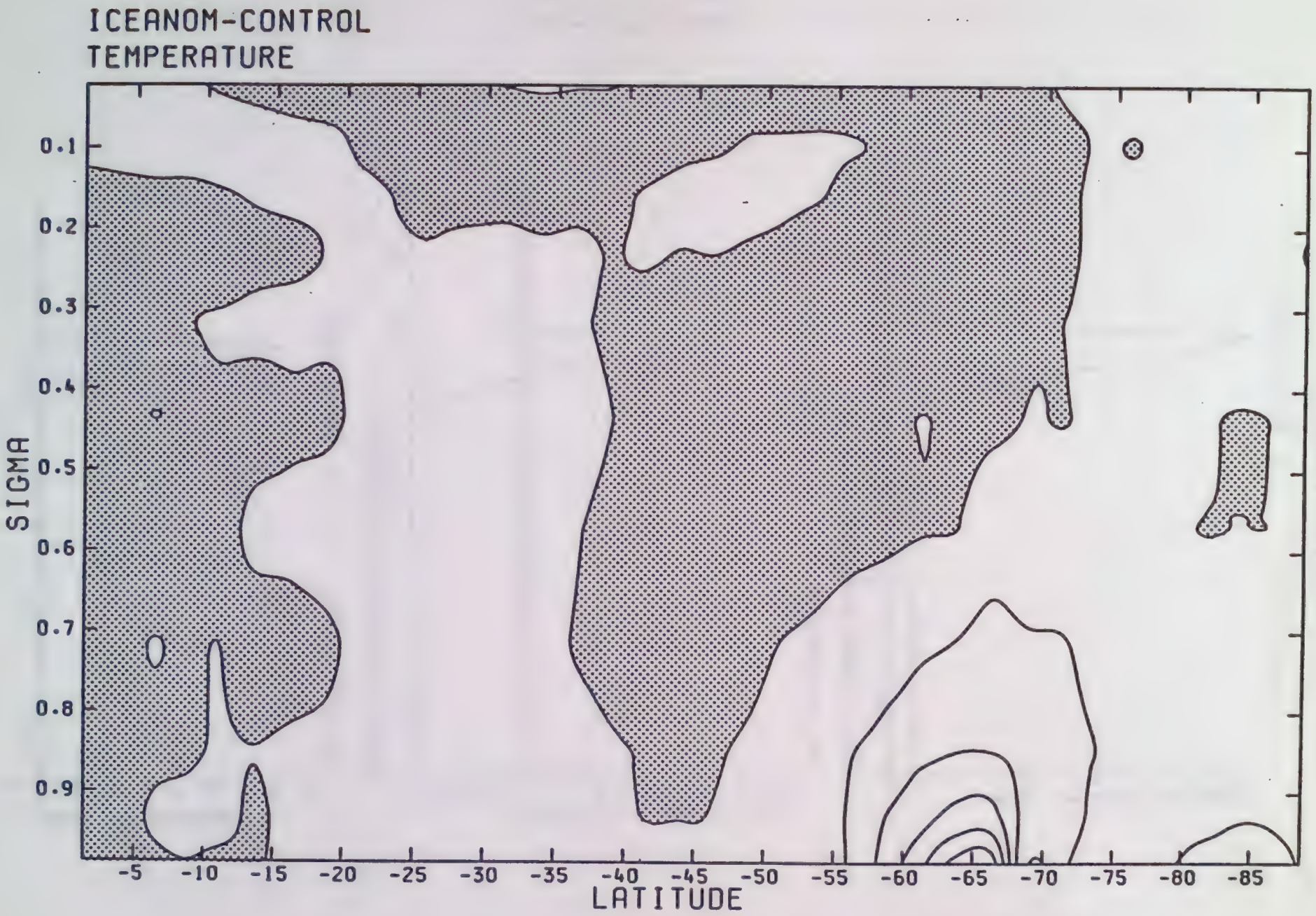

ZONALLY AVERAGED LATENT HEAT DIFFERENCES
ICEANOM-CONTROL
SMOOTH-CONTROL
LATENT HEAT DIFFERENCES WM-2
LATITUDE

FIGURE 9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TEMPERATURE
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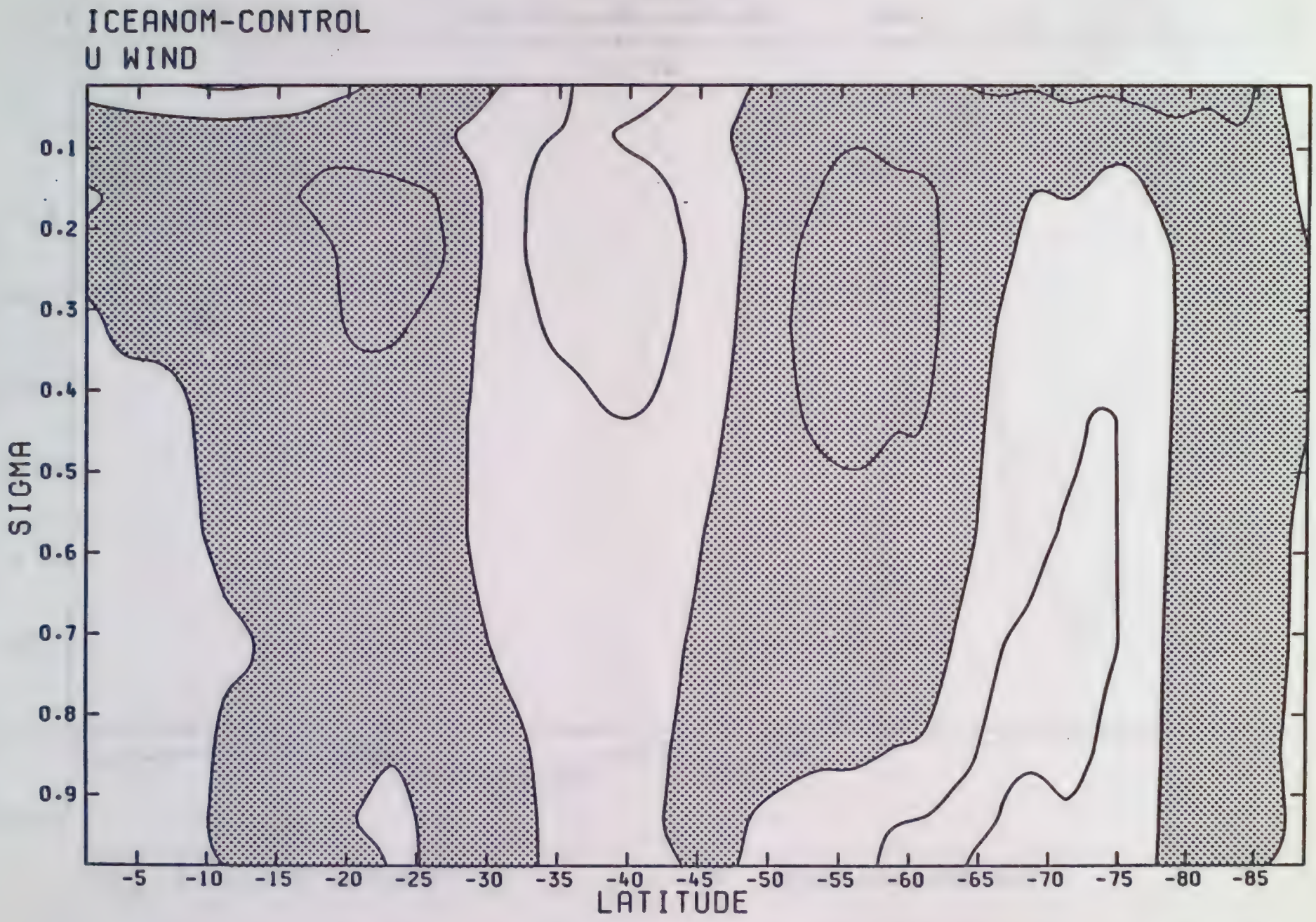
LATITUDE

FIGURE 10(a)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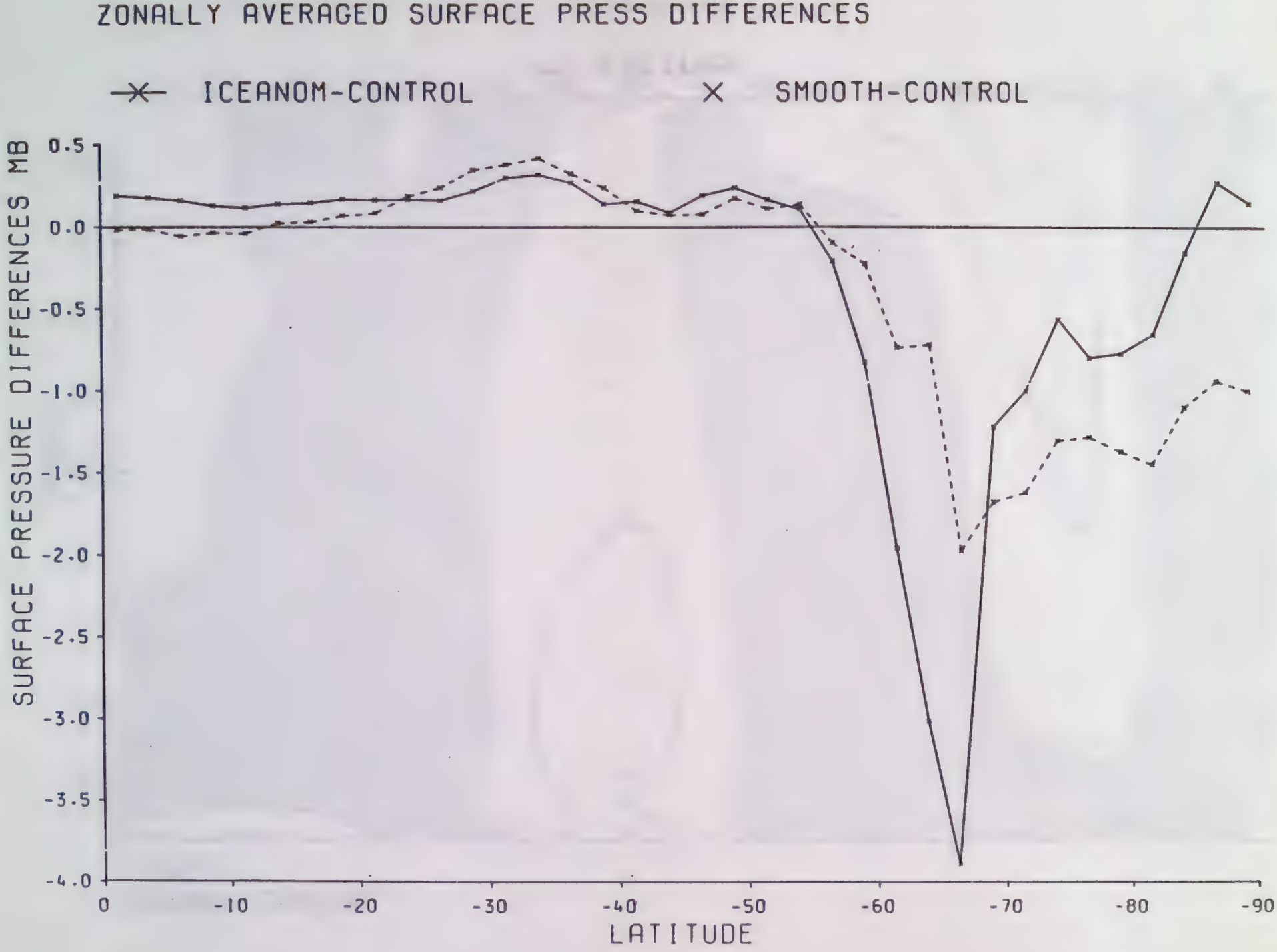
LATITUDE

FIGURE 10(b)


ZONALLY AVERAGED SURFACE PRESS DIFFERENCES
ICEANOM-CONTROL
SMOOTH-CONTROL
SURFACE PRESSURE DIFFERENCES MB
0.5
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-1.5
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LATITUDE

FIGURE 11

DIFFERENCE MAP AND T-STATISTICS WITH 6 DOF
ICEANOM RELATIVE TO 5 CONTROL JAS MEANS SURFACE PRESSURE

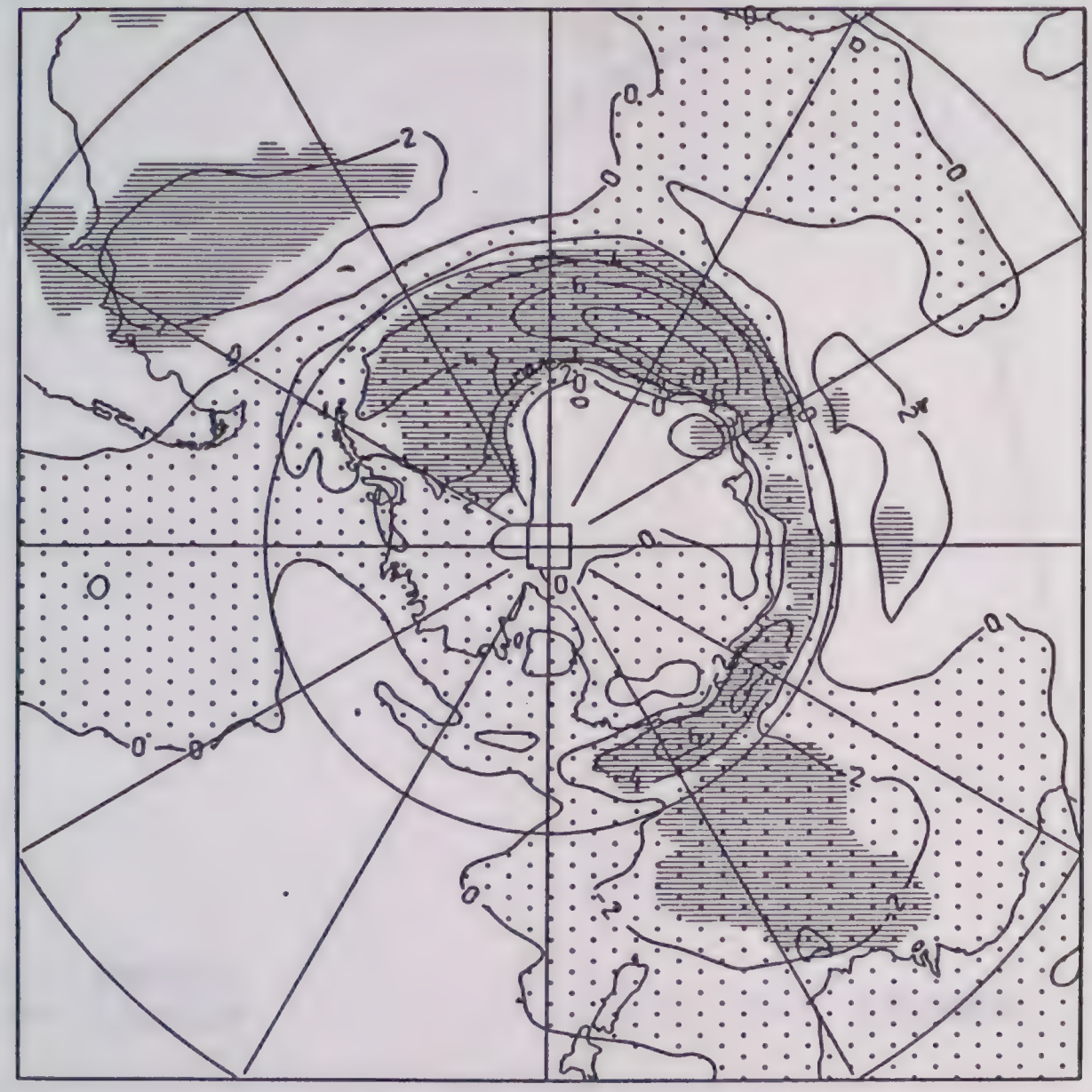
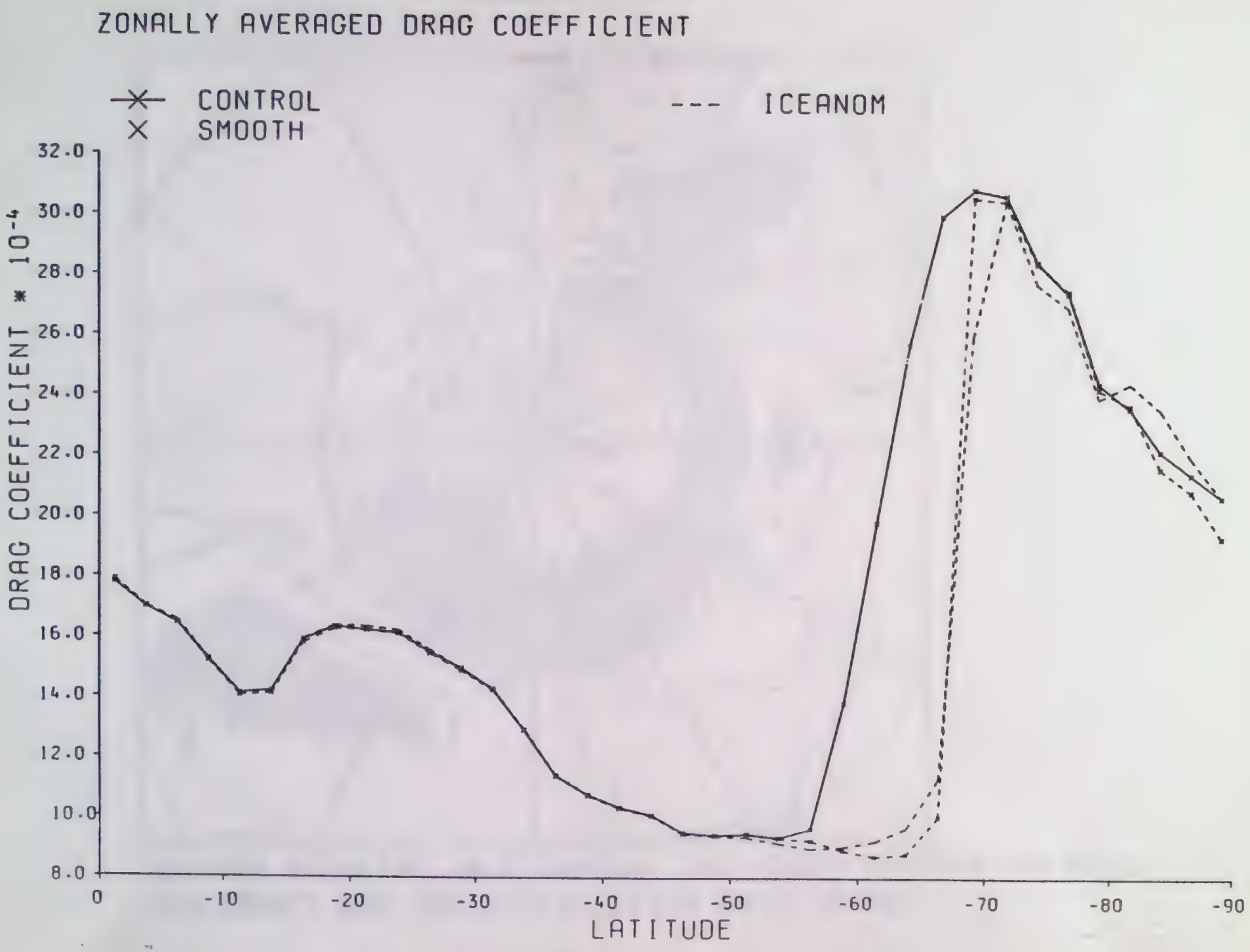


FIGURE 12


ZONALLY AVERAGED DRAG COEFFICIENT
CONTROL
SMOOTH
ICEANOM
DRAG COEFFICIENT * 10-4
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30.0
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20.0
18.0
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12.0
10.0
8.0
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-10
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-30
-40
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-60
-70
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LATITUDE

FIGURE 13

DIFFERENCE MAP AND T-STATISTICS WITH 6 DOF
SMOOTH RELATIVE TO 5 CONTROL JAS MEANS SURFACE PRESSURE

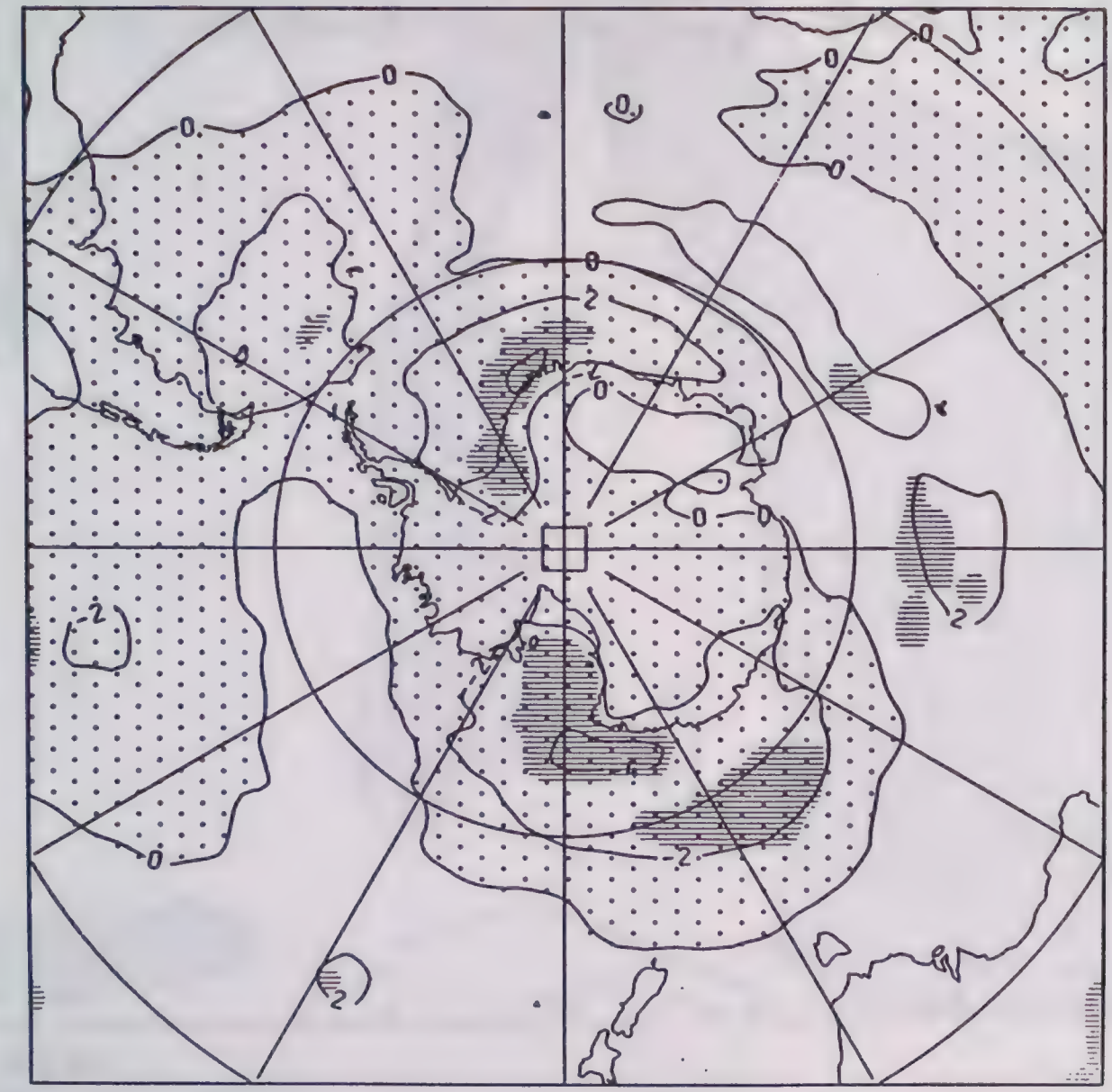
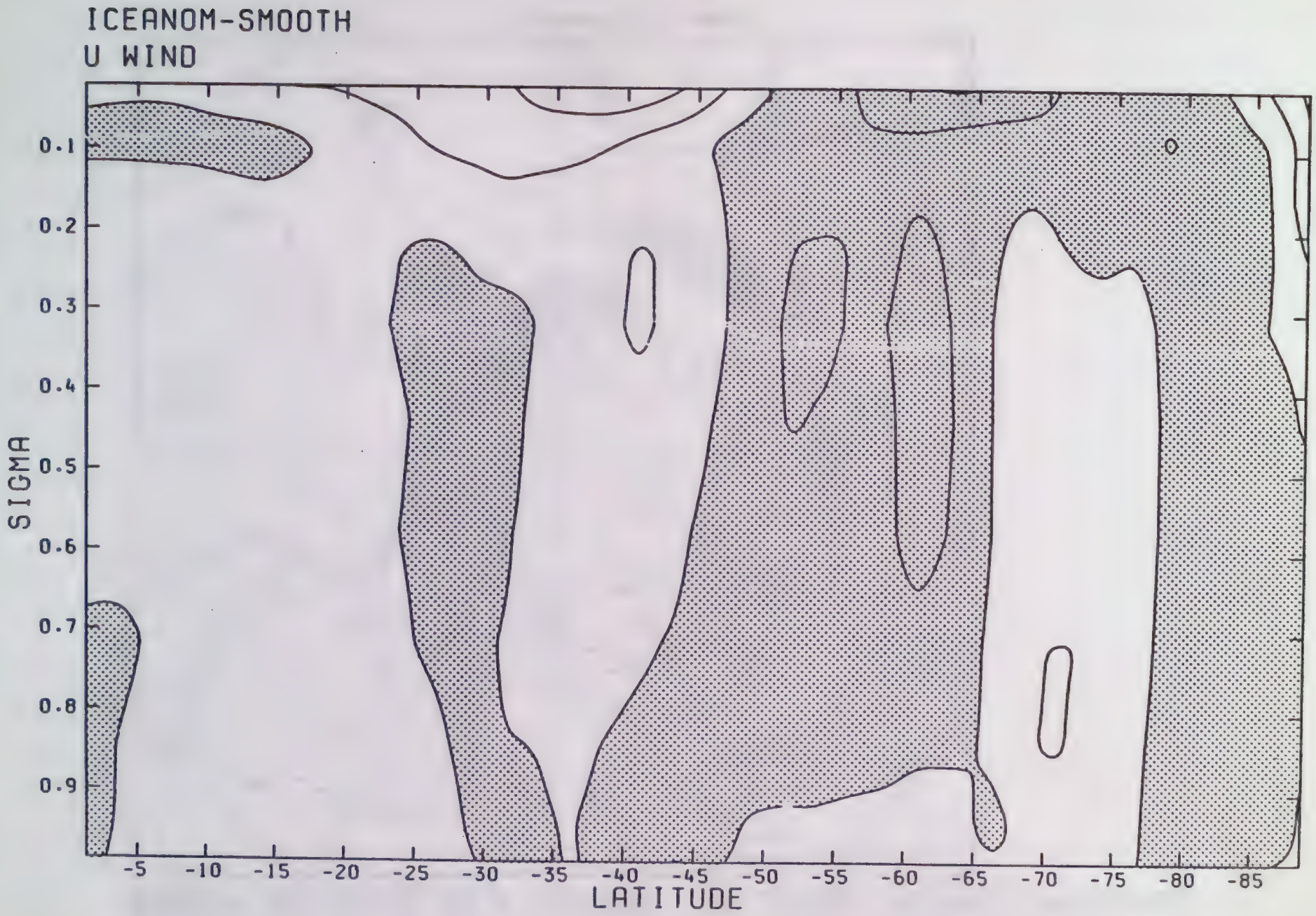


FIGURE 14


ICEANOM-SMOOTH
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-45
-50
-55
-60
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-70
-75
-80
-85
LATITUDE

FIGURE 15

DIFFERENCE MAP AND T-STATISTICS WITH 4 DOF
ICEANOM RELATIVE TO 3 SMOOTH JAS MEANS SURFACE PRESSURE

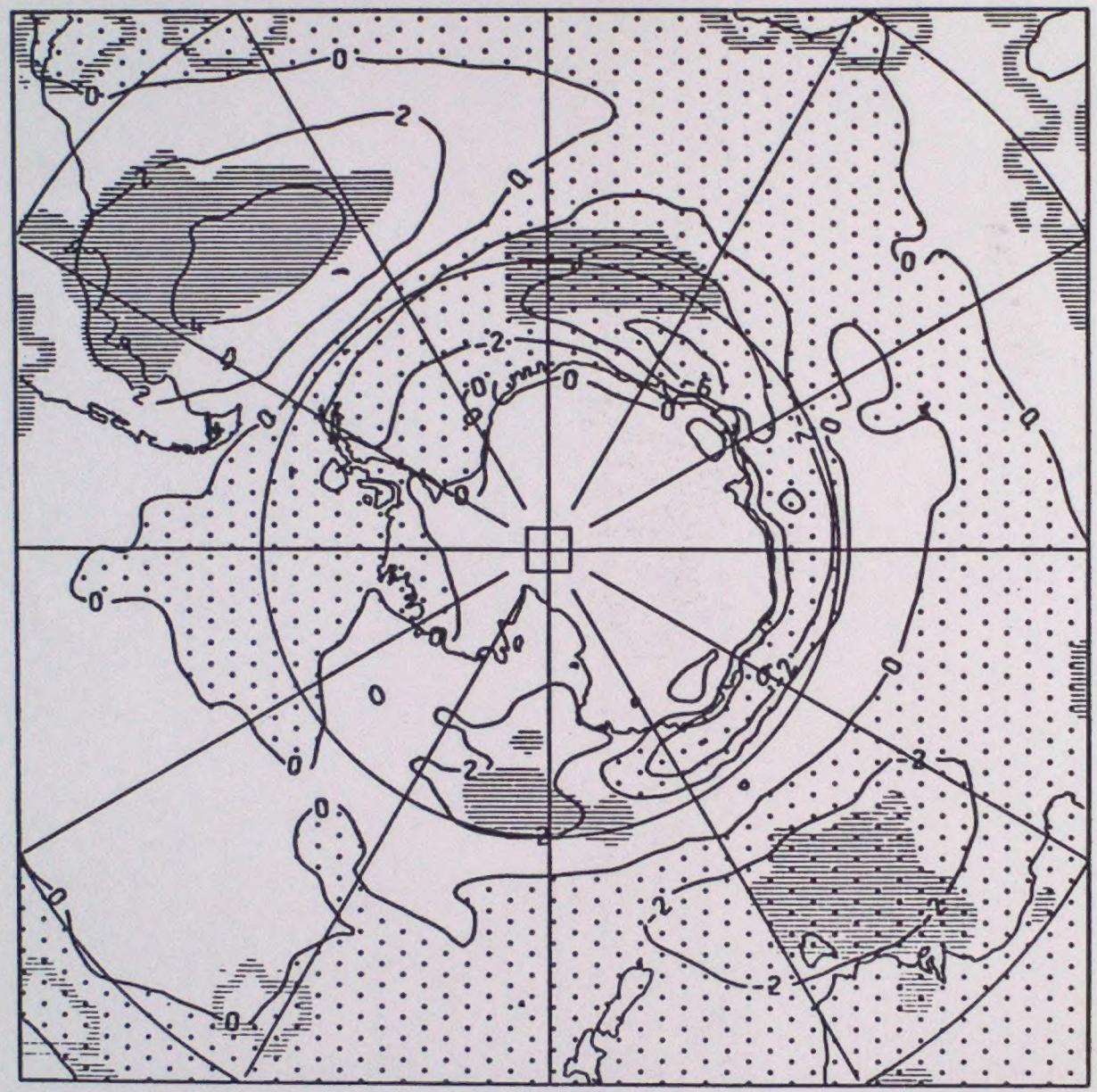

FIGURE 16

DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 1 June 1984
A scheme for incorporating the thermodynamic sea-ice model into a coupled ocean-atmosphere model.
C Gordon and M Bottomley

DCTN 2 Sept 1984
The parametrisation of the upper ocean mixed layer in coupled ocean/atmosphere models.
C Gordon and M Bottomley

DCTN 3 Sept 1984
The response of a simple thermodynamic sea-ice model to GCM forcing appropriate for perennial sea-ice.
M Bottomley

DCTN 4 Sept 1984
Potential vorticity in the stratosphere derived using data from satellites.
S A Clough, M S Grahame and A O'Neill

DCTN 5 Sept 1984
Simulation of the seasonal cycle of SST using globally specified climatological data to force simple ocean models.
M S Grahame

DCTN 6 Sept 1984
Development of an improved GCM radiation scheme.
A Slingo and R C Wilderspin

DCTN 7 Sept 1984
On the specification of surface fluxes in coupled atmosphere/ocean general circulation models.
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DCTN 8 Sept 1984
A survey of possible repercussions of the Soviet river reversal programme on Arctic sea-ice.
M Cattle

DCTN 9 Oct 1984
A study of a strong stratospheric warming during January 1982
T D A Fairlie

DCTN 10 Nov 1984
An intercomparison of the Met O 11 M45 model and the Met O 20 11-level model on the Cyber 205
J F Dyson

DCTN 11 Nov 1984
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